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PREFACE

The Journal of Marine Research is intended particularly as an outlet for shorter articles on any aspect of scientific marine investigations or any subject with a direct or indirect bearing upon the problems of the sea. Due to the intimate connection and mutual relationship between hydrospheric and atmospheric events in nature, and to the virtual identity of dynamic methods and principles in Meteorology and Oceanography, contributions to physical meteorological theory in general and to our knowledge about the relationship between ocean and air, in particular will be considered equally pertinent to marine as to meteorological research, and therefore within the scope of this Journal.

Since other facilities for the publication of the purely descriptive records and accounts of new discoveries are comparatively adequately provided for already, generally under the auspices of the institution for which the material is collected, this new journal has particularly been intended to provide a badly needed, uniform, common outlet and free forum for the discussion of new contributions to the interpretation and understanding of the sea in all its internal and external relationships, rather than to provide simply an addition to the facilities already available for placing the new factual records at the disposal of the scientific world. In the admission of contributions to this Journal, explicit theoretical contents, or general theoretical interest implicit in the nature of the recorded data *per se*, will therefore be emphasized over

cumulative additions to our factual knowledge of marine phenomena already well established in theory and empirical principle.

Although the Journal is particularly meant to fulfill the needs of North American marine and allied research, its pages will be open to all contributors. For practical reasons, publication will only be made in the English language, but aid in translation may be provided within reasonable possibility for articles accepted in other languages.

While space will be given to the discussion of any current theory in marine research in the form of signed contributions, bibliographic reviews of the literature will not be a feature of this journal, since this need is already well taken care of in other periodicals of international scope.

ON THE EVAPORATION FROM THE OCEANS

By

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In a preceding paper (Sverdrup, 1936) the author attempted a computation of the evaporation from the sea on the basis of the assumptions that:

1. A boundary layer exists next to the surface within which transport of water vapor takes place by ordinary diffusion.

2. Above the boundary layer the transport of water vapor takes place by eddy conductivity.

3. The eddy conductivity, A, is, according to laboratory results, a linear function of the altitude above the sea and depends upon the roughness of the surface which is described by means of the roughness parameter, z_o, having the dimension of a length:

$$A = \varsigma l \sqrt{\frac{\tau}{\varsigma}} = \varsigma k_o (z + z_o) \sqrt{\frac{\tau}{\varsigma}}$$

where τ is the tangential stress of the wind, ς is the density of the air, and k_o is a numerical constant equal to 0.38. The importance of the roughness is much greater than directly apparent from the above equation, since the tangential stress, τ, is a function of the roughness. The stress corresponding to a given wind velocity, measured at a distance, z, from the surface, increases with increasing roughness, and the eddy conductivity corresponding to a given velocity increases, therefore, with increasing roughness.

When undertaking a numerical computation it was furthermore assumed that the roughness parameter of the sea surface increased with increasing wind velocity and that the thickness of the boundary layer next to the surface was independent of the wind velocity. In a recent paper Rossby (1936) has discussed the roughness parameter which must be ascribed to the sea surface in order to account for the observed variation of wind velocity with altitude in the layers close to the sea surface, the observed angle between the surface and the gradient wind, and the ratio between surface wind and gradient wind. Rossby was primarily interested in the frictional force between air and water. He summarizes his results as follows:

> "In conclusion, it appears that the interaction between air and water takes place partly through real tangential shearing stresses and partly through normal pressures. With light winds and fairly smooth water,

the normal pressures appear to be insignificant. The wind velocity profile is then given by von Kármán's formula which implies the existence of a laminar subboundary layer next to the surface. With rough seas, normal pressures come into play. Within a distance from the surface which does not exceed the height of the waves, the velocity profile seems to be determined by the tangential forces mainly, but somewhat higher up the (logarithmic) velocity gradient is steeper, in response to the influence of normal pressures. In the case of rapidly changing conditions, the roughness may reach very high values (z_o = 20 cm), but with steady winds the sea surface appears to adjust itself in such a fashion as to permit the air to move over it in the most economical fashion. The roughness parameter corresponding to steady moderate to strong winds seems to be in the vicinity of 0.6 cm."

In view of these conclusions, which are based on a great amount of evidence, it is necessary to revise the previous computations of the evaporation. Instead of introducing a roughness parameter which increases with increasing wind velocity one must introduce a constant value, z_o = 0.6 cm, or consider the sea surface as being smooth. It is essential to know the correct alternative, since the value of A depends upon the tangential stress which is a function of the character of the surface. The thickness of the boundary layer through which vapor transport takes place by diffusion must be studied, since the assumption of a constant thickness appears improbable because over a smooth surface the thickness of the laminar sub-layer is nearly inversely proportional to the wind velocity.

Before proceeding it is necessary to consider the mechanism of the transport of water vapor. In the case of a rough surface above which no layer of laminar flow exists, the eddy conductivity has a definite value at the surface itself, meaning that the eddies actually reach to the surface. In spite of this, it is probable that when dealing with the transport of water vapor one must introduce a boundary layer within which the transport of water vapor takes place through diffusion. One can conceive that for a short time a mass of air comes to rest in one of the hollows of the surface. This mass of air will immediately give off its momentum to the surface, but the vapor pressure will not immediately attain the value which is characteristic of the surface. As long as this mass remains at rest its vapor contents will be increased by processes of diffusion, supposing that it originally was lower than the equilibrium pressure corresponding to the temperature and salinity of the surface. After a short time the air mass will be removed and replaced by another, and similar processes will be repeated. In order to describe these processes one can introduce a boundary layer the thickness of which is a statistical quantity representative of the average conditions and within which the transport of water vapor takes place through diffusion.

TABLE 1

MEAN VALUES OF VAPOR PRESSURE AT DIFFERENT LEVELS ACCORDING TO OBSERVATIONS ON BOARD THE ATLANTIS

Group	*n*	u_6 *m/sec*	t_4 *mb*	e_6 *mb*	e_s *mb*	e_1 *mb*	$e_{1.5}$ *mb*	e_2 *mb*	e_4 *mb*	e_8 *mb*	e_{18} *mb*	e_{38} *mb*	*c*
b_1	5	2.4	23.7	24.68	28.24	25.14		24.99	24.91			24.12	0.286
a_1	9	2.6	22.6	24.17	28.09		24.63		24.30			23.54	0.336
b_2	3	2.7	21.6	14.18	28.00	14.93		14.63	14.45			13.37	0.419
a_2	4	2.9	22.4	17.03	30.80		17.48		17.28			16.31	0.374
c_1	3	3.1	24.8	25.38	31.90			25.90	25.40	25.33		24.60	0.419
d	3	3.4	26.1	25.72	34.27			26.40	25.80	25.65	25.13		0.545
g	6	4.5	22.2	16.77	30.12			17.48	17.18			15.37	0.740
f	6	4.6	21.4	16.32	26.87		17.21		16.68	16.08		15.12	0.655
a_3	4	4.65	24.3	21.32	32.25		22.52		21.50			19.84	0.812
c_2	8	5.3	24.2	19.75	31.29			20.83	20.21	19.46		17.77	1.038
e	8	5.3	24.5	23.59	31.74				23.71	23.63		22.24	0.695
h	10	6.6	19.9	15.78	24.47				16.06			14.52	0.685

If, on the other hand, a layer of laminar flow existed, and if there were an abrupt transition between the pure laminar and the turbulent regions, the thickness of the laminar sub-layer would, according to von Kármán (1934), be given by $d = 11.5\nu/(\tau/\varsigma)^{\frac{1}{2}}$, where ν is the kinematic coefficient of friction. Von Kármán points out, however, that the transition is gradual and that the value $d = 30\nu/(\tau/\varsigma)^{\frac{1}{2}}$ appears as a more reasonable limit for the region influenced by laminar friction. Under these conditions it can be assumed that the transport of water vapor by diffusion takes place where the motion is laminar, and that the thickness of the diffusion layer lies between the stated limits.

If the eddy conductivity within the turbulent region is known, the thickness of the sub-layer of diffusion can be computed by means of observations of the vapor pressure at different levels above the sea. When the first computation was attempted only one single series of such observations was

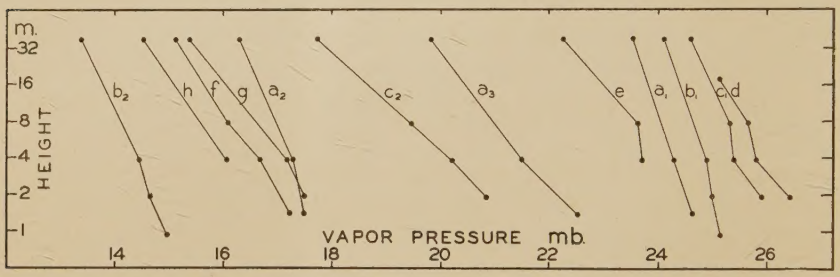

Figure 1. Vapor pressure plotted against the logarithm of altitude according to observations on board the ATLANTIS.

available (Wüst, 1920), but recently Mr. R. B. Montgomery made a number of such observations on board the ATLANTIS, and kindly placed them at my disposal. Table 1 contains mean values based on the latter series in the cases in which the potential temperature decreased with height. The measurements could not always be undertaken at the same levels, but when computing mean values only observations from the same levels were combined in groups. Within each group the observations were arranged according to the wind velocity, and in the table the mean values are entered in the order of increasing wind velocity. This was measured at an altitude of 6 meters. The interpolated value of the vapor pressure at 6 meters is also entered.

In figure 1 the mean values of the vapor pressure are plotted against the logarithm of altitude. It is seen that in most cases the values fall on straight lines and that the deviations as a rule are of a magnitude which must be expected, since errors of observations enter with considerable weight owing to the small number of series in each group.

It appears, thus, to be well established that the vapor pressure is a linear function of the logarithm of height, if the potential temperature decreases with height, that is, if instability exists. Wüst (1937) has recently pointed out that according to a detailed analysis of his data of 1920, the logarithmic law is not valid below a level of about 50 cm, but this fact is of minor importance since we are now dealing with processes at some distance from the sea surface. Within this region the validity of the logarithmic law has recently been confirmed by special observations undertaken by Lieutenant F. L. Black on board the U. S. S. Indianapolis.

Supposing that the upward transport of water vapor is the same at all levels, it follows that the eddy conductivity is a linear function of height:

$$A = a(z + z_o)$$

As previously, we assume

$$A = \varsigma k_o \sqrt{\frac{\tau}{\varsigma}}(z + z_o) = \varsigma k_o w(z + z_o), \tag{1}$$

where, according to von Kármán, $w = \sqrt{\frac{\tau}{\varsigma}}$ is called the friction velocity.

The value of z_o and the value of w, corresponding to a measured velocity at some distance from the surface, depend upon the character of the surface.

Within the turbulent region the transport of water vapor, F, in gr/cm^2 sec is:

$$F = -\frac{df}{dz}A = -\frac{0.623}{p}\frac{de}{dz}A \tag{2}$$

where f is the specific humidity of the air, $f = \frac{0.623}{p - e}e$, and e is the vapor pressure. Since the vapor pressure is a logarithmic function of height, we can write

$$e = e_o - cln\frac{z + z_o}{z_o} \quad \text{or} \quad e = e_d - cln\frac{z + z_o}{d + z_o} \tag{3}$$

For each of the groups in Table 1, the value of c has been computed and entered in the last column of the table.

We obtain, therefore,

$$F = \frac{0.623}{p}\varsigma k_o wc \tag{4}$$

The transport of water vapor through the layer of diffusion is, on the other hand,

$$F = \frac{\delta}{d}(e_w - e_d) \tag{5}$$

where δ is the coefficient of diffusion referred to vapor pressure, d is the thickness of the layer of diffusion and where e_w and e_d represent the vapor pressure at the surface and at the upper boundary of the diffusion layer, respectively. Since

$$\delta = \kappa \frac{0.623}{p} \varsigma \tag{6}$$

where κ is the coefficient of diffusion in cm²/sec (0.235), we obtain from (3), (4), and (5):

$$d = \frac{\kappa}{k_o w}\left(\frac{e_w - e}{c} - ln\frac{z + z_o}{d + z_o}\right) \tag{7}$$

The values of z_o and w depend, as already stated, upon the character of the surface, and the values of F and d will, therefore, depend upon our assumptions concerning the surface.

In the case of a rough surface, we have

$$z_o = \text{roughness parameter} \tag{8}$$

$$w = \sqrt{\frac{\bar{\tau}}{\varsigma}} = \frac{k_o}{ln\frac{z + z_o}{z_o}} u_z = \frac{0.165}{\log\frac{z + z_o}{z_o}} u_z \tag{9}$$

where u_z is the velocity at the level z.

In the case of a smooth surface we assume

$$z_o = \frac{\nu}{k_o w} - d \tag{10}$$

since for $z = d$ we must have $A/\varsigma = \nu$. The value of w can be computed from the equation (v. Kármán, 1934):

$$\frac{u}{w} + 5.57 \log\frac{u}{w} = 5.5 + 5.75 \log\frac{zu}{\nu} \tag{11}$$

Rossby (1936) has published a curve showing u/w as a function zu/ν, and this curve will be used here.

By means of the above formulae w, F, and d have been computed from the data in Table 1, assuming either that the sea surface can be considered as a rough surface with a roughness parameter of $z_o = 0.6$ cm, or that the sea surface can be considered as a smooth surface. The computed values are shown in Table 2 in which also the ratio $F/(e_w - e_6)$ has been entered. It is seen that the results from groups b_2 and a_2 deviate much from the

others and these groups will, therefore, be left out of consideration. It is possible that in these cases stationary conditions were not reached, for which reason our computation gives fictitious values, or that the roughness of the sea surface was much greater than usual, which, according to Rossby, may occur.

Assuming the surface to be rough and characterized by $z_o = 0.6$ cm, we find nearly twice as great values of the evaporation as when assuming the surface to be smooth. It should be observed that this result which is based on equations (4), (9), and (11) is derived completely from consideration of the vapor transport in the turbulent layer and is independent of any assump-

TABLE 2

VALUES OF THE FRICTION VELOCITY, w, THE EVAPORATION, F, AND THE THICKNESS OF THE LAYER OF DIFFUSION, d, ASSUMING ROUGH OR SMOOTH SURFACE

	Rough surface, $z_o = 0.6$ cm.				*Smooth surface*			
Group	w	10^6F	$\frac{10^6F}{e_w - e_6}$	d	w	10^6F	$\frac{10^6F}{e_w - e_6}$	d
	cm/sec	*gr/cm²sec*		*cm*	*cm/sec*	*gr/cm²sec*		*cm*
b_1	13.2	1.06	0.30	0.28	7.6	0.61	0.17	.22
a_1	14.3	1.34	0.34	0.22	8.2	0.77	0.20	.15
b_2	14.8	1.74	0.13	1.14	8.5	0.98	0.07	1.53
a_2	15.9	1.66	0.12	1.26	9.1	0.95	0.07	1.80
c_1	17.0	1.98	0.30	0.33	9.7	1.13	0.17	.34
d	18.7	2.86	0.33	0.31	10.6	1.62	0.19	.31
g	24.8	5.15	0.39	0.29	13.8	2.86	0.21	.33
f	25.3	4.64	0.45	0.23	14.1	2.59	0.25	.24
a_3	25.6	5.82	0.53	0.16	14.2	3.24	0.30	.13
c_2	29.2	8.49	0.74	0.09	15.9	4.63	0.40	.03
e	29.2	5.68	0.70	0.10	15.9	3.09	0.38	.05
h	36.3	6.98	0.80	0.10	19.6	3.75	0.43	.07

tion concerning the thickness of the diffusion layer. Since we do not know the amount of evaporation when the measurements were undertaken, we have no possibility of deciding which assumption concerning the character of the surface is correct, but the values of d may give a clue.

The values of d decrease with increasing wind velocity. In figures 2 and 3 they are plotted against $1/w$. Assuming the surface to be rough, we find that d is nearly proportional to $1/w$, and taking the number of series within each group into account, we obtain

$$d = \frac{4.12}{w} = \frac{27.5\nu}{w} \tag{12}$$

The values of d, which have been computed on the assumption of a smooth surface, show such wide scattering that representation by means of

Figure 2. Thickness of the diffusion layer plotted against $1/w$ (w = friction velocity), assuming the sea surface to have a roughness parameter of 0.6 cm.

Figure 3. Thickness of the diffusion layer plotted against $1/w$ (w = friction velocity) assuming the sea surface to be smooth.

a straight line appears to be arbitrary, but introducing a straight line we obtain

$$d = \frac{1.86}{w} = \frac{12.4\nu}{w} \tag{13}$$

The last value is of the magnitude which should be expected from von Kármán's theory. Our analysis, therefore, does not give a result which helps in deciding whether we, when computing the evaporation, have to consider the sea surface as rough or smooth. In order to arrive at a decision it is necessary to undertake a comparison between computed and observed values of the evaporation.

Evaporation values obtained from measurements in pans on board ships must be reduced in order to represent the evaporation from the sea surface. Such a reduction was undertaken by Wüst (1920), but he probably arrived at too low values since he found that the average evaporation from the ocean is 82 cm per year. A later revision gave 93 cm, but H. Mosby (1936) found 106 cm when considering the energy available for evaporation. Assuming 100 cm as a probable value we must, therefore, multiply Wüst's values of 1920 by 1.22.

Wüst has published average values of the evaporation for regions of the Atlantic Ocean and has compiled the average wind velocities from these regions such that w can be computed, but the average humidity gradients are not known, for which reason formula (4) cannot be used for a computation of the evaporation. From (3), (4), and (5), however, we obtain

$$F = \frac{\beta(e_w - e_z)w}{\frac{\kappa}{k_o} ln \frac{z + z_o}{d + z_o} + dw} \tag{14}$$

Introducing d from (12) or (13) we can, therefore, compute the evaporation if we know the wind velocity at one level, by means of which w is found, the temperature and relative humidity at one level, by means of which e_z is found, and the surface temperature and salinity, from which e_w is derived. Such values are available from the Atlantic Ocean between latitudes 50° N and 55° S. These and the computed and observed values of the evaporation expressed in mm per 24 hours are entered in Table 3. The regions are the same as those treated by Wüst. The temperatures of the air and the sea surface are derived from Schott's "Geographie des Atlantischen Ozeans," the relative humidities are based on Wüst's information, and the mean wind velocities are taken from Wüst's paper. It has been assumed that all observations have been taken at an altitude of 6 meters above the sea surface. The vapor pressure at the surface, e_w, is obtained by multiplying the vapor pressure corresponding to the surface temperature by 0.98 in order to take the effect of the salinity into account.

TABLE 3

METEOROLOGICAL DATA AND SURFACE TEMPERATURES FROM REGIONS OF THE ATLANTIC OCEAN
COMPUTED AND OBSERVED VALUES OF THE EVAPORATION FOR THE SAME REGIONS

Region	*Temperature, °C.*		*Rel. humidity*	e_w	e	$e_w - e$	u	*Evaporation in 24 hours*		
	Water	*Air*	*%*	*mb*	*mb*	*mb*	*m/sec*	E_s	E_r	*Obsv'd**
50° to 40° N	10.8	10.5	82	12.74	10.10	2.64	8.4	1.2	2.3	2.2
40 to 30	18.3	17.2	80	20.64	15.73	4.91	6.7	1.8	3.3	3.2
30 to 8	25.4	24.9	76	31.82	24.92	7.90	6.7	2.8	5.2	4.3
8 to 3	27.4	26.8	83	35.80	29.26	6.55	2.8	1.1	1.9	3.0
3 N to 20 S	25.8	25.7	78	32.62	25.80	6.82	6.1	2.2	4.1	4.0
20 S to 40 S	19.5	18.3	80	22.25	16.85.	5.40	5.6	1.7	3.2	3.2
40 to 55	9.9	8.7	82	12.00	9.26	2.74	7.8	1.1	2.2	1.6

* According to Wüst (1920) but multiplied by the factor 1.22.

In figure 4 the observed and computed values of the evaporation are presented graphically. The computed values which are based on the assumption of a rough surface agree nearly with the observed values, but those based on the assumption of a smooth surface and a layer of laminar flow next to the surface give values which are about half of the observed ones. It appears, therefore, that the evaporation from the ocean can be

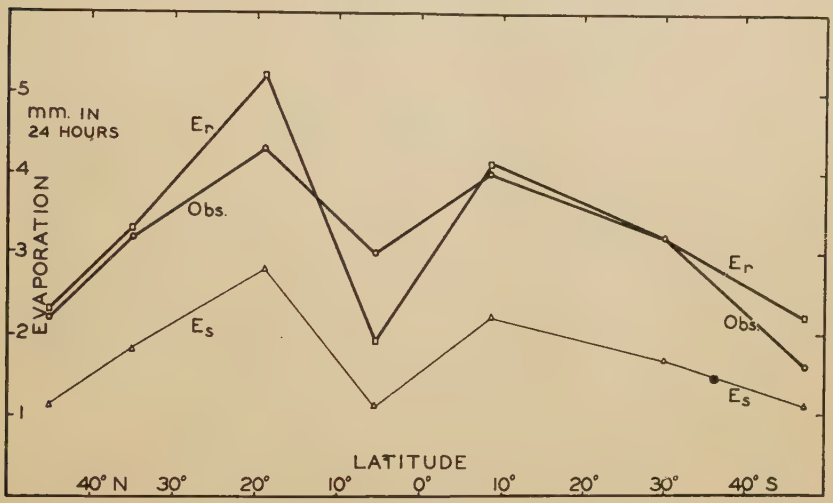

Figure 4. Observed values of evaporation over the Atlantic oceans, and computed values, based on the assumption that the sea surface is characterized by a constant roughness parameter (E_r) or that it is a smooth surface (E_s).

computed on the basis of the present theory, provided (1) that the sea surface can be considered as a rough surface characterized by the roughness parameter, $z_o = 0.6$ cm; and (2) that a layer of diffusion of thickness, $d = 27.5\nu/w = 4.12/w$, exists next to the surface. The first assumption is in agreement with Rossby's conclusions concerning the roughness of the sea surface, but both this and especially the second assumption should be studied by means of additional data. If such studies confirm our results, equation (14) may perhaps be applied when studying the amount of water vapor which an airmass receives when flowing over the ocean.

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ON THE MUTUAL ADJUSTMENT OF PRESSURE AND VELOCITY DISTRIBUTIONS IN CERTAIN SIMPLE CURRENT SYSTEMS*

By

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The writer (*1*) recently advanced the hypothesis that the horizontal pressure gradients observed in the current systems of the atmosphere and the ocean to a large extent must be interpreted as reactions to the Coriolis' forces impressed upon these systems by the rotation of the earth. Observational support for this point of view was found in the fact that both the temperature-salinity and oxygen-salinity correlation curves obtained from stations on both sides of the Gulf Stream in the region between Nova Scotia and Bermuda are very nearly identical, even though the individual isotherms may drop as much as 700 meters between the slope water basin and the Sargasso Sea. This mass distribution and the resulting pressure distribution are most readily interpreted as the result of a continuous banking process caused by the action of the Coriolis' forces on the moving masses of the Gulf Stream system. Theoretical support was found in a theorem by Taylor (*2*), according to which any purely two-dimensional motion which can occur in a non-rotating system is equally possible also when the system rotates at uniform speed around an axis normal to the plane of motion, the deflecting forces due to the rotation then being offset by the reaction of the fluid system in the form of a superimposed "Coriolian" pressure field. It is easily shown that this nullification of the Coriolis' forces is possible only when the prescribed motion is purely two-dimensional.

Both in the atmosphere and in the ocean balancing pressure gradients are established through the piling up of mass in certain regions and the removal of mass from others, i. e. through vertical motions. Thus the conditions upon which the complete validity of Taylor's theorem depends are not

* This note constitutes a preliminary report on certain theoretical investigations now in progress at the Massachusetts Institute of Technology. Other phases of the problem discussed below are being investigated by Messrs. J. Holmboe, G. Grimminger and H. Wexler. The author is greatly indebted to Mr. Grimminger for all the numerical computations in connection with the evaluation of the functions F and Φ referred to below and for the preparation of the various diagrams. Mr. Grimminger's computations were in part carried out with the aid of the differential analyzer belonging to the Electrical Engineering Department of the Massachusetts Institute of Technology; the assistance of the staff members in charge is gratefully acknowledged.

satisfied during the time required for the establishment of the "Coriolian" pressure field. For this reason it seems desirable to study, in some detail, the process whereby dynamic balance is created and to determine if there are any limitations to the ability of a given system to build up a Coriolian pressure distribution. The existence of such limitations would necessitate modifications in the current theory previously suggested by the author on the basis of Taylor's theorem.

There are some indications that such limitations exist. Experiments with wake streams in rotating tanks carried out by Mr. Spilhaus (*3*) and described elsewhere in this issue suggest that complete balance between pressure gradient and deflecting force is never reached. In all cases examined by Mr. Spilhaus the current pattern indicates an inadequate development of the Coriolian pressure field, resulting in anticyclonic current trajectories resembling inertia circles.

Recently attempts have been made at the Massachusetts Institute of Technology to determine trajectories in the atmosphere through a study of the movements of tongues of moist and dry air within selected isentropic surfaces (*4*). The daily charts constructed for this purpose show that individual air currents have a pronounced tendency to break up into large anticyclonic eddies, presumably as a result of inadequate development of the Coriolian pressure gradients.

To the author's knowledge no theory exists which satisfactorily describes the mechanism whereby the mass and pressure distributions adjust themselves to the velocity distribution, although the problem is of great practical importance; its solution would be of value not only to physical oceanographers but also to meteorologists in connection with the interpretation of the so-called dynamic pressure formations (i. e. warm anticyclones, cold cyclones). The problem is now being analyzed in some detail by the author and his collaborators. The purpose of the present article is merely to call attention to the important rôle played by lateral (isentropic) mixing in this process.

To illustrate the mechanism of adjustment it is advisable to analyze first the changes which must occur in a straight parallel current encircling the earth at a given latitude and characterized by complete symmetry with respect to the axis of the earth. A somewhat simpler system may be obtained by considering the effect of lateral diffusion on a linear current system in a rotating plane. It is in this form that the problem will be studied below.

We shall first describe the lateral diffusion of momentum in a parallel current in a homogeneous, incompressible and non-rotating medium of constant density ρ. A mathematical theory for such a current system is easily obtained through an adaptation of Tollmien's experimentally well established theory for two dimensional wake streams (*5*). Fig. 5 demon-

strates three successive velocity profiles resulting from lateral diffusion within such a system. If one designates the velocity by u and measures the horizontal y-coordinates to the left from the down-stream axis of the current, the equation of motion reduces to

$$\rho D \frac{\partial u}{\partial t} = \frac{\partial \tau D}{\partial y}, \tag{1}$$

where t is the time and D the depth of the fluid, which in this case remains constant from point to point and throughout the entire process. There

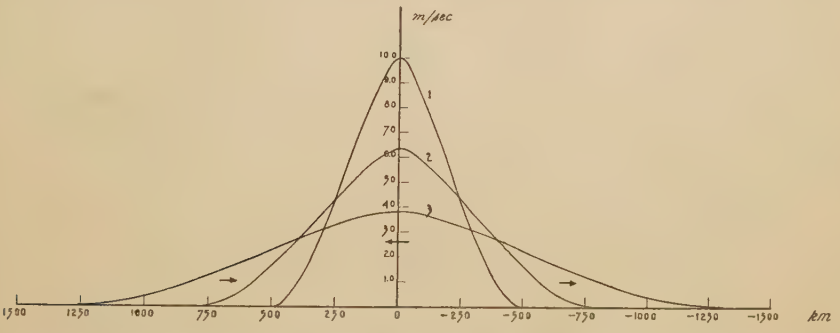

Figure 5. Successive velocity profiles in stationary system. Arrows indicate direction of transversal displacements which would result from the action of a superimposed Coriolis force.

are no transversal velocity components. The shear stress τ vanishes in the resting fluid surrounding the current and thus

$$\rho D \int_{-\infty}^{+\infty} u dy = \text{constant}. \tag{2}$$

The integral (2) measures the momentum of the current and remains constant throughout the diffusion process. Thus the areas under the various curves in Fig. 5 are equal.

Proceeding now to the diffusion of momentum in a rotating system it is evident that equation (1) and the velocity profiles in Fig. 5 should represent the results of the diffusion process also in a rotating system, provided transversal displacements do not appear and provided the shearing stresses are unaffected by the rotation. If it be assumed that the time rate of diffusion is so slow that the current at each instant remains in dynamic equilibrium (balance between pressure gradient and deflecting force), successive profiles of the free surface should then have the shape indicated in Fig. 6. These curves have been computed from the curves in Fig. 5 with the aid of the equation for gradient motion,

$$u = -\frac{g}{f}\frac{\partial D}{\partial y}, \tag{3}$$

in which expression g represents the acceleration of gravity and f the Coriolis' parameter.

An inspection of Fig. 6 reveals that the change in pressure distribution illustrated by these curves cannot be accomplished without some transversal transfer of mass (northward, in case the current moves eastward). As soon as such transversal displacements occur, equation (1) no longer ex-

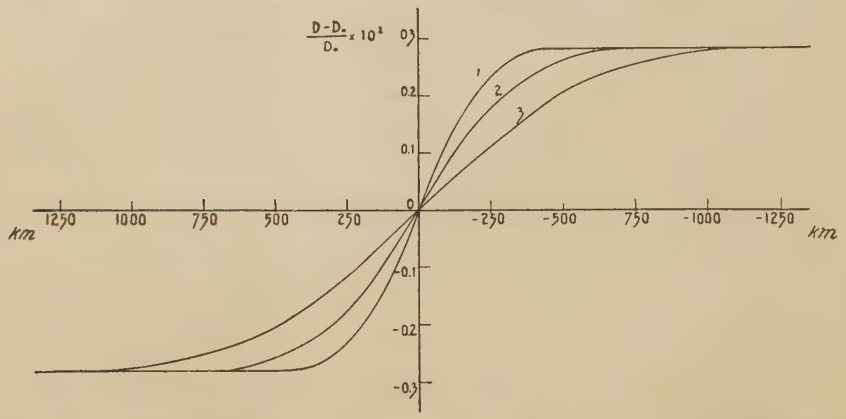

Figure 6. Successive free surface profiles corresponding to the velocity profiles in figure 5, assuming at all times equilibrium between Coriolis' force and pressure gradient. The curves indicate that the gradual adjustment of the free surface can not take place without transfer of mass from right to left across the current. The value of D_0 used in computing these curves from the ones in Figure 5 corresponds to the height of the homogeneous atmosphere (8 km.).

presses the true balance of forces. Without resort to formulae, the general character of the diffusion process may be described as follows:

Through lateral mixing the layers immediately left and right of the main current will receive a slight acceleration down-stream while the center portion will lose in speed. Thus the immediate effect is to establish velocities in excess of the gradient value to the left and right of the current axis and of less than the gradient value in the axis itself. The resulting unbalance produces slight transversal velocities towards the left in the center of the current while the immediate environment, characterized by supergradient velocities, tends to move towards the right. These transversal components, which are indicated by the arrows in Fig. 15, must produce convergence with pressure rise in the left half of the current system, divergence with pressure drop in the right; the mass distribution is thus gradually adjusted towards the new velocity distribution.

Since the transversal velocities must vanish at great distances from the axis it is evident that the fluid portions far to the left of the current must experience divergence, those far to the right convergence. Thus the current is gradually set off from its environment by a trough of low pressure to the left, a ridge of high pressure to the right.

A quantitative estimate of the pressure changes described above may be obtained with the aid of the momentum integral. The equations of motion and the equation of continuity for the linear current system here described may be written in the following form:

$$\rho D \frac{du}{dt} = \rho f v D + \frac{\partial \tau D}{\partial y} \tag{4}$$

$$\rho D \frac{dv}{dt} = - \rho f u D - \rho g D \frac{\partial D}{\partial y} \tag{5}$$

$$\frac{\partial D}{\partial t} + \frac{\partial v D}{\partial y} = 0 \tag{6}$$

In these equations axial velocities are designated by u, transversal velocities by v.

With the aid of the equation of continuity (6), the first equation of motion may be written in the form

$$\frac{\partial}{\partial t}[\rho D(u - fy)] + \frac{\partial}{\partial y}[\rho v D(u - fy)] = \frac{\partial \tau D}{\partial y}. \tag{7}$$

Integration across the current system into the undisturbed surrounding regions gives

$$\int_{-\infty}^{+\infty} \frac{\partial}{\partial t}[\rho D(u - fy)]\, dy = 0 \tag{8}$$

or

$$\frac{\partial}{dt} \int_{-\infty}^{+\infty} u D dy = f \int_{-\infty}^{+\infty} y \frac{\partial D}{\partial t}\, dy. \tag{9}$$

Equation (8) expresses the fact that the diffusion process leaves the total *absolute* momentum of the current system unchanged.

If the diffusion is slow it may be assumed that the current system passes through a series of successive dynamic equilibrium states; consequently the second equation of motion (5) reduces to the form

$$\frac{\partial D}{\partial y} = - \frac{fu}{g}. \tag{10}$$

The integral on the left side of (9) may then be integrated, giving

$$-\frac{g}{f}\frac{\partial}{\partial t}[D_{\infty}{}^{2} - D_{-\infty}{}^{2}] = f\int_{-\infty}^{+\infty} y\frac{\partial D}{\partial t}dy. \tag{11}$$

Since the depth of the undisturbed fluid to the right ($D_{-\infty}$) and to the left ($D_{+\infty}$) of the current do not change, it follows that in this case of slow diffusion

$$\int_{-\infty}^{+\infty} y\frac{\partial D}{\partial t}dy = 0 \tag{12}$$

It is evident that the curves in fig. 6 do not satisfy this requirement. The establishment of the successive profiles in fig. 16 would require a net transfer of fluid to the left and this transfer must, under the influence of the deflecting force, lead to the development of a net axial component in excess of the gradient value.

The magnitude of this net component (δu) is readily computed. If the final boundaries of the current are given by $y = \pm a$ it may be shown that the excess velocity component has a mean value over the entire current cross section from $-a$ to $+a$ of

$$\delta u = \frac{f\int_{-a}^{+a} y(D - D_i)dy}{\int_{-a}^{+a} Ddy}, \tag{13}$$

in which D_i represents the initial depth and $D - D_i$ thus measures the change in depth at each point. The immediate effect of such an unbalanced velocity component is to remove fluid from the left edge of the current system and to pile it up along the right edge, in such a fashion that a trough of low pressure and a ridge of high pressure are formed, as described above.

In reality the process of diffusion and the adjustment of the mass distribution must take place continuously and not in the discrete steps described in the previous analysis. To follow the continuous diffusion process it is necessary to integrate the equations of motion (4, 5, 6). This will now be done under the assumption that the diffusion process is so slow that the system passes through a series of dynamic equilibrium states. Under those conditions the equations reduce to the simpler form

$$\rho fvD + \frac{\partial \tau D}{\partial y} = 0 \tag{14}$$

$$fu + g\frac{\partial D}{\partial y} = 0 \tag{15}$$

$$\frac{\partial D}{\partial t} + \frac{\partial vD}{\partial y} = 0. \tag{16}$$

It is easily seen that the first of these equations satisfies the momentum integral

$$\int_{-\infty}^{+\infty} y \frac{\partial D}{\partial t}\, dy = 0, \tag{17}$$

derived previously for the case of dynamic equilibrium. From a combination of the first and the third equation it follows that

$$-\rho f \frac{\partial D}{\partial t} + \frac{\partial^2 \tau D}{\partial y^2} = 0. \tag{18}$$

This equation may be integrated provided we know the value of the shearing stress. Let ν represent the kinematic eddy viscosity characteristic of the lateral mixing process. Then

$$\frac{\tau}{\rho} = \nu \frac{\partial u}{\partial y} = -\frac{\nu g}{f} \frac{\partial^2 D}{\partial y^2}, \tag{19}$$

and consequently

$$\frac{\partial D}{\partial t} + \frac{g}{f^2} \frac{\partial^2}{\partial y^2}\left(\nu D \frac{\partial^2 D}{\partial y^2}\right) = 0. \tag{20}$$

If one introduces for νD a constant value $\nu_o D_o$*, where D_o represents the mean depth of the fluid, equation (20) reduces to the form

$$\frac{\partial D}{\partial t} + \nu_o \cdot \frac{g D_o}{f^2} \frac{\partial^4 D}{\partial y^4} = 0. \tag{21}$$

This equation may be written in a simpler form if one introduces non-dimensional time and length scales,

$$y = \lambda \eta \qquad \left(\lambda = \frac{1}{f}\sqrt{g D_o}\right) \tag{22}$$

$$s = \frac{\nu_o t}{\lambda^2} \tag{23}$$

Then

$$\frac{\partial D}{\partial s} + \frac{\partial^4 D}{\partial \eta^4} = 0 \tag{24}$$

and consequently also

* The assumption $\nu D = \nu_o D_o$ is obviously quite satisfactory in the case of a deep homogeneous medium. In case of a stratified medium this assumption implies that within a given layer the kinematic eddy viscosity resulting from lateral mixing is inversely proportional to the depth of that particular isopycnic (in the atmosphere, isentropic) layer. This assumption is in qualitative agreement with Parr's (*6*) results concerning the relation between vertical stability and the intensity of isopycnic (isentropic) mixing.

$$\frac{\partial u}{\partial s} + \frac{\partial^4 u}{\partial \eta^4} = 0. \tag{25}$$

This final equation is easily integrated. If one assumes that the initial velocity distribution is given by

$$u_o = \psi(\eta), \tag{26}$$

the velocity distribution at any subsequent time s will be given by

$$u = \frac{1}{\pi}\int_{-\infty}^{+\infty} \psi(\eta + \alpha\sqrt[4]{4s})\, F(\alpha)\, d\alpha, \tag{27}$$

*

the function $F(\alpha)$ being defined by the integral

$$F(\alpha) = \int_0^\infty e^{-\frac{\lambda^4}{4}} \cos \lambda\alpha \cdot d\lambda, \quad \int_{-\infty}^{+\infty} F(\alpha)\, d\alpha = \pi. \tag{28}$$

As an illustration we shall compute the gradual diffusion of a current characterized initially by a total pressure drop cross-stream of the amount $\rho g\Delta$. Thus the initial depth of the fluid is $D_o - \frac{1}{2}\Delta$ to the left of the current and $D_o + \frac{1}{2}\Delta$ to the right of the current. It is not difficult to formulate the solution for the case of a finite initial width of the current. However, the initial width of the current is of little significance in the later stages of the diffusion process. For this reason, and in order to gain simplicity, it will be assumed that the initial width is infinitely small. Thus, from (27) one finds

$$D = \frac{1}{\pi}\int_{-\infty}^{+\infty} \psi(\eta + \alpha\sqrt[4]{4s})\, F(\alpha)\, d\alpha, \tag{29}$$

where

$$\psi(\eta) = D_o - \frac{\Delta}{2} \qquad \text{for } \eta > 0. \tag{30}$$

and

$$\psi(\eta) = D_o + \frac{\Delta}{2} \qquad \text{for } \eta < 0. \tag{31}$$

Thus

$$\psi(\eta + \alpha\sqrt[4]{4s}) = D_o - \frac{\Delta}{2} \quad \text{for } \alpha > -\frac{\eta}{\sqrt[4]{4s}} \tag{32}$$

and

$$\psi(\eta + \alpha\sqrt[4]{4s}) = D_o + \frac{\Delta}{2} \quad \text{for } \alpha < -\frac{\eta}{\sqrt[4]{4s}} \tag{33}$$

* A detailed discussion of equation (25) and its solution (27) will be published later.

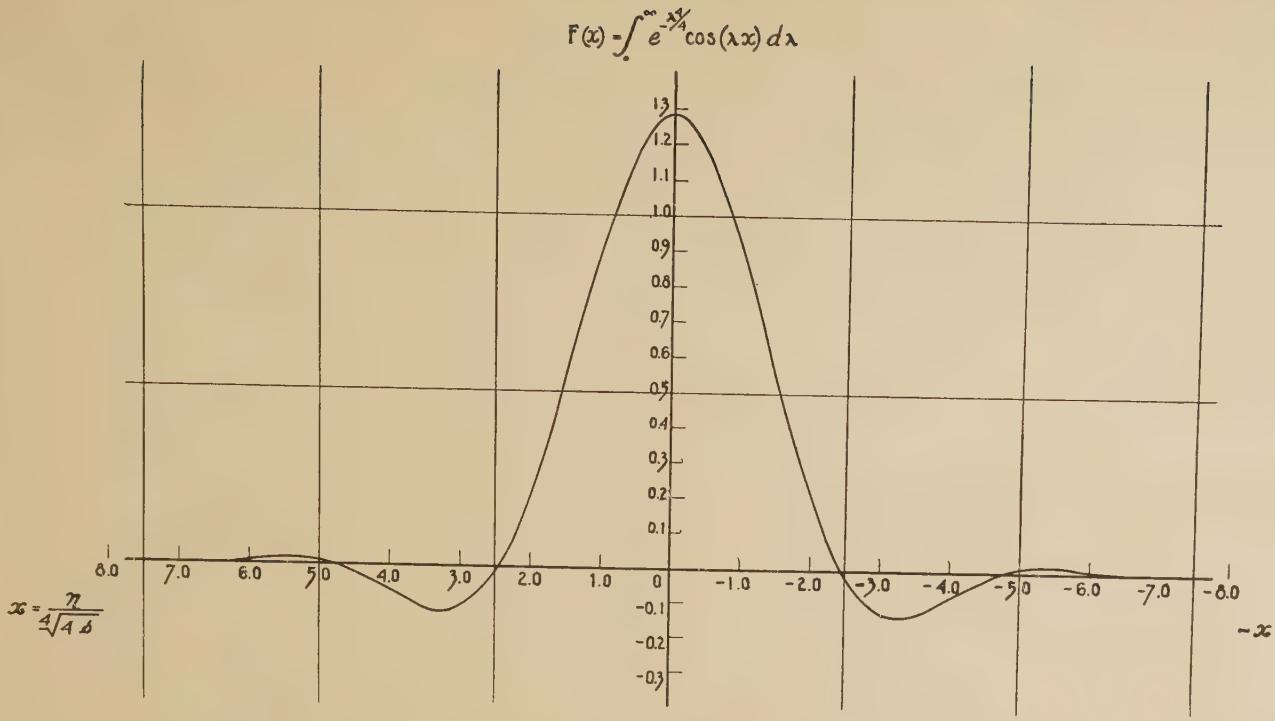

Figure 7. Non-dimensional representation of velocity distribution as a function of $\frac{\eta}{\sqrt[4]{4s}}$, taking into consideration the Coriolis' force acting upon the transversal velocity components. The ordinate is 1.2818 u/u_o, where u_o is the simultaneous velocity in the center of the current.

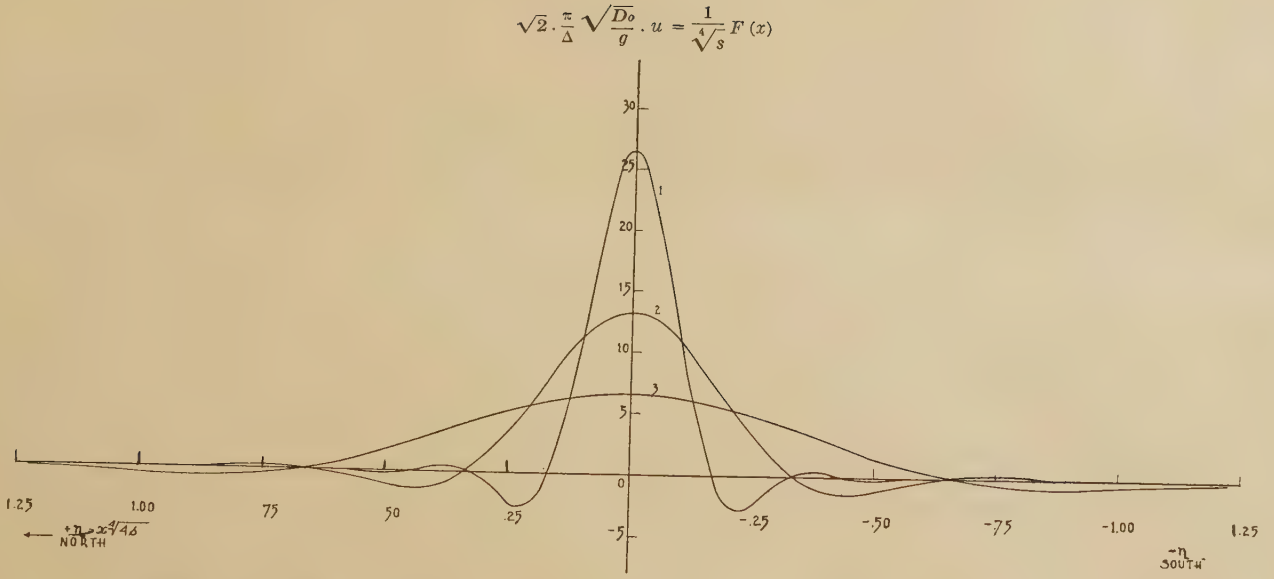

Figure 8. Successive velocity profiles, computed for $s_1 = 0.055 \cdot 10^{-4}$, $s_2 = 16\ s_1$, $s_3 = 16\ s_2$, and plotted against the non-dimensional transversal coordinate η. To obtain actual velocities from the non-dimensional ordinates plotted in the figure, the latter must be multiplied with the constant factor $\frac{\Delta}{\pi} \sqrt{\frac{g}{D_0}} \cdot \frac{1}{\sqrt{2}}$.

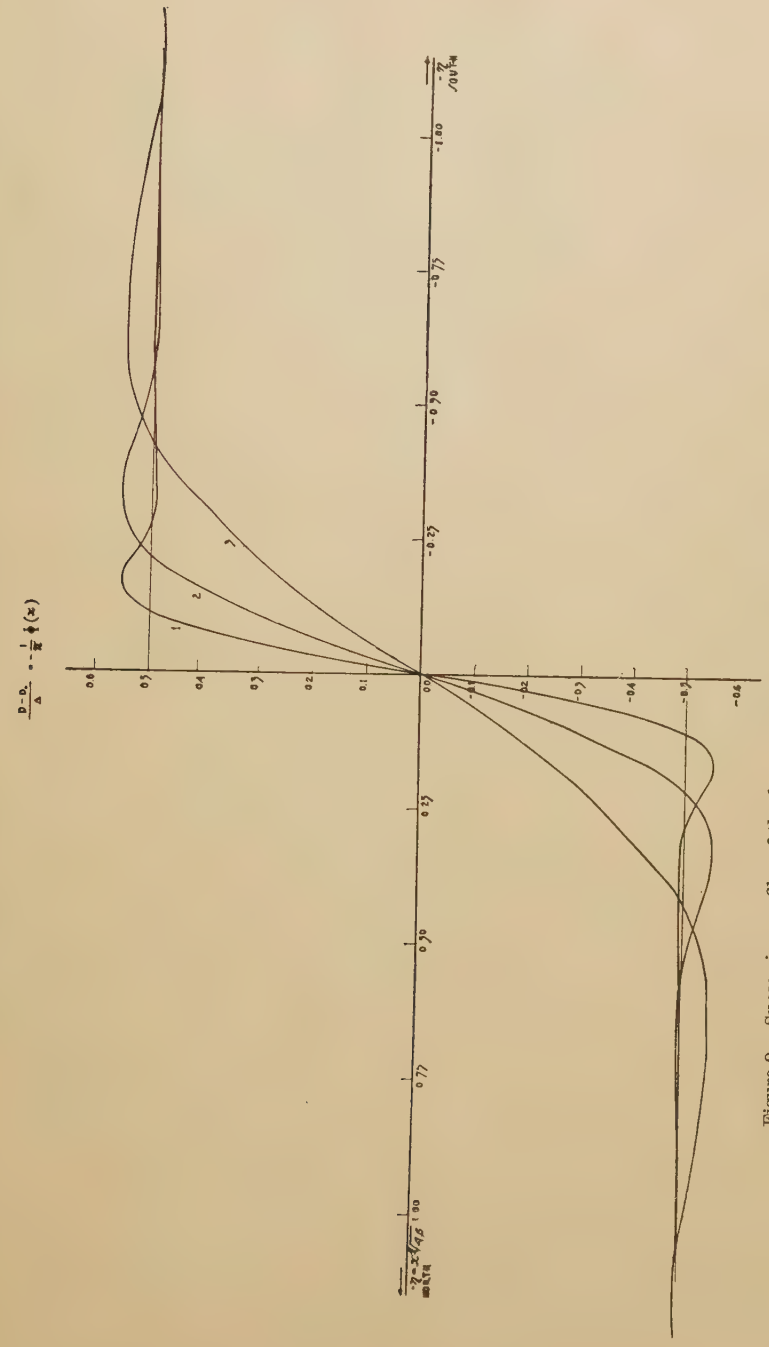

Figure 9. Successive profiles of the free surface, corresponding to the velocity profiles in Figure 8.

It follows that

$$(34)\quad D = \frac{1}{\pi}\left(D_o - \frac{\Delta}{2}\right)\int_{-\frac{\eta}{\sqrt[4]{4s}}}^{\infty} F(\alpha)\,d\alpha + \frac{1}{\pi}\left(D_o + \frac{\Delta}{2}\right)\int_{-\infty}^{-\frac{\eta}{\sqrt[4]{4s}}} F(\alpha)\,d\alpha.$$

After reductions and in view of the fact that $F(\alpha)$ is an even function one finds

$$(35)\quad D = D_o - \frac{\Delta}{\pi}\int_0^{\frac{\eta}{\sqrt[4]{4s}}} F(\alpha)\,d\alpha$$

or

$$(36)\quad D = D_o - \frac{\Delta}{\pi}\Phi\left(\frac{\eta}{\sqrt[4]{4s}}\right),$$

where

$$(37)\quad \Phi(r) = \int_0^r F(\alpha)\,d\alpha$$

The successive velocity profiles are computed from the gradient wind equation, which may be written in the form

$$(38)\quad u = -\frac{g}{f\lambda}\frac{\partial D}{\partial \eta} = -\sqrt{\frac{g}{D_o}}\frac{\partial D}{\partial \eta}$$

or

$$(39)\quad u = \frac{\Delta}{\pi}\sqrt{\frac{g}{D_o}}\frac{1}{\sqrt[4]{4s}}F\left(\frac{\eta}{\sqrt[4]{4s}}\right).$$

The velocity distribution is represented in non-dimensional form in fig. 7, this being a plot of

$$(40)\quad F_o \cdot \frac{u}{u_{\eta=o}} \equiv F\left(\frac{\eta}{\sqrt[4]{4s}}\right)$$

against $\frac{\eta}{\sqrt[4]{4s}}$. A conventional representation of the velocity distribution ψ at three consecutive times is given in Fig. 8. Fig. 9 contains a representation of the free surface, giving the distribution of

$$(41)\quad \frac{1}{\pi}\Phi\left(\frac{\eta}{\sqrt[4]{4s}}\right) = -\frac{(D - D_o)}{\Delta}$$

against η for a few selected values of s.

These diagrams demonstrate that the diffusion leads to the development of surrounding counter currents, having a maximum intensity of about eleven per cent of the simultaneous maximum velocity in the axis of the current. From the depth profiles it is seen that the total drop in pressure across the current increases through the development of a trough to the left, a ridge to the right. The percentual increase in the total pressure drop across the current is about eight percent. The spread of the current, as measured by the distance from trough to ridge is proportional to $s^{-\frac{1}{2}}$, i. e. at first rapid, then increasingly slow.

Preliminary estimates indicate that the banking process described above becomes more intense when the effect of stratification is taken into account, but this phase of the problem has not yet been sufficiently well investigated to warrant discussion at the present time.

SUMMARY

Recent investigations suggest that the current systems of the atmosphere and the ocean have a definite tendency to break up in large-scale anticyclonic eddies. This result points towards a limitation in the ability of these current systems to build up, through the accumulation or removal of mass, such compensating horizontal pressure gradients as would be required to offset completely the Coriolis' forces resulting from the rotation of the earth. As a first step in an investigation aiming at the determination of these limitations and the consequent breaking up of the currents, a study has been made of the changes in the mass distribution which accompany the lateral diffusion of momentum in a straight parallel current on the surface of the earth. It is found that the diffusion process is attended by weak transversal velocity components having a net component from high towards low pressure. Through the action of the Coriolis' force associated with this mass transfer the current will at all times have a slight net component downstream in excess of the gradient value. The Coriolis' force corresponding to this excess axial velocity produces a banking of the fluid towards the right of the current axis and thus the formation of a trough of low pressure along the left edge of the current, a ridge of high pressure along the right edge. The magnitude of the pressure changes thus created are determined for the case of very slow diffusion and are found to be about 8% of the total initial pressure drop across the current.

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NOTE ON THE FLOW OF STREAMS IN A ROTATING SYSTEM

By

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Certain similarity of the observed flow in the ocean and in the atmosphere to the behavior of a wake stream, for which a complete, experimentally verified theory is available (*1*), led Rossby (*2*) to initiate the extension of the wake stream idea first to a rotating system and secondly, to a stratified medium. Several interesting predictions were made by Rossby as a result of his theoretical investigation; and it was the purpose of the experiments carried out by the author to investigate to what extent these predictions could be verified in the laboratory. The experiments were of an exploratory nature and the conclusions can in no way be regarded as final.

Preliminary tests were made in a small cylindrical tank subject to cyclonic rotation and with a jet so arranged that the influx of water took place radially from the periphery. With this system the path described by water injected through a point jet at the surface was demonstrated for various radii of the inertia circle and these flow patterns were compared with those obtained in the same fashion but with the jet extending the whole depth of the layer of water being investigated. In the latter case the flow patterns differed radically from the inertia circle, the anticyclones being very much larger for the deep jet than for the surface, point jet. This indicated that certain pressure gradients were established which served to balance to some extent the Coriolis' forces. That such pressure gradients will be established has been demonstrated mathematically by Taylor (*3*).

It was evident, however, that with the jet issuing from the side of the tank, the wall effect contributed to the balancing pressure gradients. To minimize this wall effect, therefore, a large tank 6 feet in diameter was constructed and the jet inflow arranged to take place radially outwards from the center. The water was brought to the tank from a stationary head cistern by means of a gland around the hollow axis of the tank and issued through a vertical slit in this hollow axis. In this system it was possible to vary the following parameters at will:—

(I) the depth, from 5 cm. to 40 cms.
(II) the head at the jet, from 10 cms. to 100 cms.
(III) the rotational speed, from 1¾ r.p.m. to 10 r.p.m.

* Contribution No. 158

From Taylor's theory there is reason to expect, as indicated above, that the Coriolis or deflecting forces due to rotation would be completely balanced by pressure gradients resulting from the motion. Therefore, in a deep homogeneous medium with two dimensional motion, the flow pattern should be the same regardless of rotation and in the cylindrical tank described would take the form of a cyclone on the left side of the jet looking downstream and a symmetrically placed anticyclone on the right. Although Taylor's conditions were in no sense satisfied in the experimental arrangement described, they are approached more closely as the depth of the water in the tank increases, and therefore the flow patterns should show greater symmetry in the deeper cases. This was found to be the case; though, again, even in the deepest test made, the pressure gradients were far from completely balancing the deflecting force.

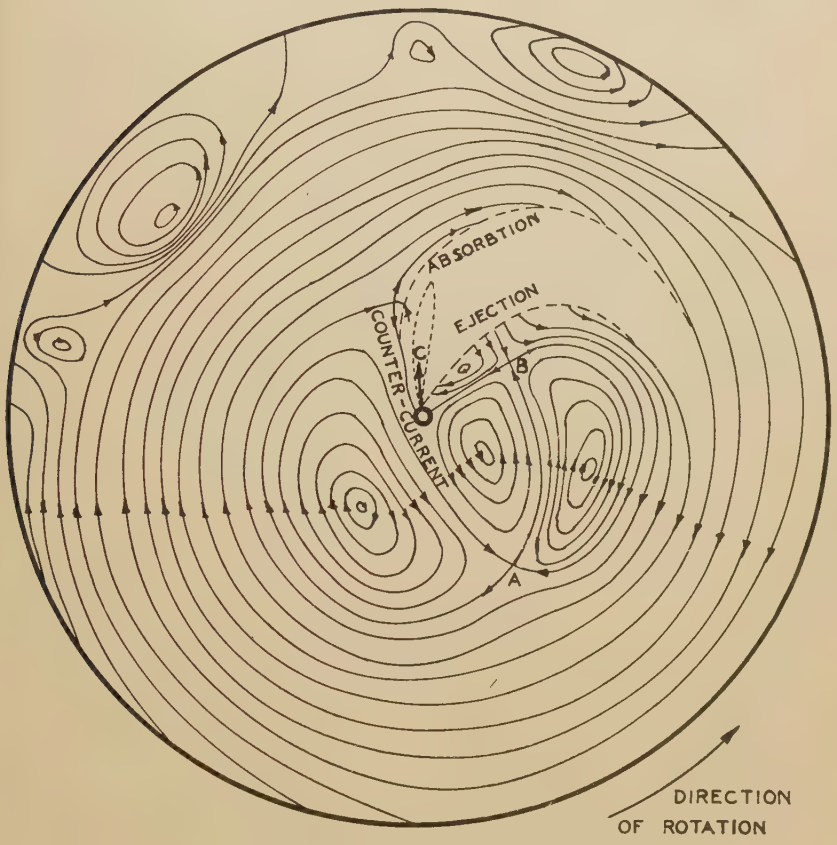
Rossby's extension of the wake stream theory to a rotating system calls for a counter-current on the left hand side of the jet, due to absorption there and ejection on the right side. This counter-current is brought about, in a homogeneous fluid, by deformations of the free surface and should be expected to be most marked for shallow depths. Thus it is seen that, whereas the counter-current does not appear to be compatible with the conception of completely balanced Coriolis forces, the two results are actually complementary cases at the extremes of the depth scale, viz.—in shallow layers the deformation of the free surface is percentually great compared to the depth and therefore the counter-current can be established, but in deep layers, where the surface deformations are percentually negligible, Taylor's reasoning applies.

This was well demonstrated in the six foot tank. With a head of 40 cms. and a rotation of 5 revolutions per minute, a clear counter-current accompanying a completely asymetrical flow pattern was observed for a depth of 6 cms., while for the same conditions of head and rotation, *i. e., the same radius of the inertia circle,* no counter-current and a markedly more symmetrical flow pattern was found when the depth was increased to 24 cms.

It has been pointed out that the effect of a given deformation of the free surface increases with diminishing depth, but decrease of depth also has the undesirable characteristic of accentuating the influence of bottom friction. The surface deformation effect can, however, be intensified, without decreasing the depth to the extent that bottom friction would play a noticeable role, by utilizing a stratified medium. Moreover, in nature, stratification often exists in the atmosphere and always is present in the ocean so that, while Taylor's conclusions are true for a homogeneous deep fluid, Rossby's speculations as to the motion in shallow depths and stratified media are the more significant in Meteorology and Oceanography by virtue of their closer relationship to the natural state.

In order to demonstrate this effect of stratification a rotating tank was

built having a four foot diameter and with 12 cms. depth; apart from its size this tank was similar in all respects to the larger one described above. Into the four foot tank was poured first a layer 6 cm. deep of a mixture, specific gravity 1.2, of Carbon Tetrachloride and Kerosene, and then above



KEY DIAGRAM

Figure 10. Surface flow in four foot rotating tank with stratification.

this the jet injected the wake stream into an overlying layer of water of equal depth. The water was allowed to overflow all around the edge of the tank, and when it is pointed out that the jet was only 12 thousandths of an inch wide it becomes evident that the distributed overflow would not

introduce any serious effect on the flow pattern. In the ideal case with this stratification, if the lower liquid were to rotate with the tank as a solid, deformations of the free water surface would be magnified at the internal boundary by a factor

$$\frac{\rho_{lower}}{\rho_{lower} - \rho_{upper}} = \frac{1.2}{0.2} = 6,$$

Figure 11. Surface flow in four foot rotating tank with stratification, ten revolutions per minute cyclonic rotation and 100 cms. head at jet.

and therefore the Rossby effects are magnified while the Taylor balance is diminished. In the practical experiment it is impossible to prevent the circulation of the water above from imparting a similar motion to the underlying denser liquid, and, hereby, a tendency to barotropy, which would

nullify the effectiveness of the stratification, occurs. Complete barotropy is prevented by bottom friction acting on the lower liquid and by the fact that the jet only plays in the overlying water and therefore a more intense counter-current and a less symmetrical flow is expected in the stratified than in the corresponding homogeneous case. Comparison of flow patterns for the two-layer and the homogeneous case shows this expectation to be justified.

In the case illustrated by the photograph reproduced here, chosen from an extensive series, the rotation was 10 revolutions per minute, the head 100 cms. and the two-layer stratification was employed. The flow may be easily traced by reference to the accompanying Key Diagram. Lycopodium powder was used as the indicating substance and the flow was photographed with the camera rotating with the tank. Examination of the pattern reveals that several of the features predicted are ratified to some extent at least in the experiment.

(I) A fully developed counter-current is established on the left hand side of the stream near the jet.

(II) Further downstream absorption on the left side is noticeable and such absorption implies the existence of a return current there also, but this counter-current is masked by the water sweeping around in the general anticyclonic motion and overlapping the wakestream.

(III) Ejection downstream on the right hand side is evident while on this side the only absorption noticeable occurs in the immediate proximity of the jet.

While these features to some extent are in accordance with Rossby's predictions it is also evident that important modifications to his analysis will have to made before the behavior of a wake stream in a rotating system can be completely explained.

It is interesting to note the flow at A and B in the photograph; the camera has caught two hyperbolic points in a classical pose.

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SEASONAL VARIATIONS IN THE SURFACE WATERS OF SAN JUAN CHANNEL DURING THE FIVE YEAR PERIOD, JANUARY 1931 TO DECEMBER 30, 1935

By

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INTRODUCTION

Although there is a mass of data available on the physical and chemical properties of sea water in various parts of the world, relatively little data may be had from localities where continuous long time observations have been made. At the Oceanographic Laboratories on San Juan Channel near Friday Harbor, Wash., the water body is unique in that it is rather homogeneous for a depth of 100 meters. We propose to demonstrate the yearly and seasonal variations and fluctuations of the physical and chemical properties, and the phytoplankton calendar of this water mass and to determine if long time periodicities exist.

Data on the seasonal variation of the physical and chemical properties of this water have been previously published, showing the changes over a three-year period (*13*, *21*, *22*). These data include temperature and chlorinity observations for the entire period, results of pH and dissolved oxygen for two years, and the data for phosphate and silicate concentrations for one year. The present paper deals with determinations, for the period of January 1931 to December 1935, on variations in temperature, chlorinity, dissolved oxygen, phosphate, silicate, pH, nitrates and nitrites. Included also are observations on the rainfall during the period of study. The phytoplankton calendar will appear shortly.

GEOGRAPHY AND LOCATION

The accompanying chart, Fig. 12, shows the location of the Oceanographic Laboratories in the San Juan Archipelago. General descriptions of this region have been given in previous publications (*12*, *16*, *24*).

The waters of San Juan Channel have their origin primarily in the Strait of Juan de Fuca. In the late spring and summer months, especially during periods of northerly or northwesterly winds, the surface waters are sometimes very markedly affected by waters of low densities carried in to the channel from Georgia Strait. Extensive studies have been made of the waters of the latter by investigators at the Pacific Biological Station at Nanaimo, British Columbia (*10*, *11*, *14*, *15*). The surface waters of

Georgia Strait are materially affected by the discharge of the Fraser River, and decided stratification of various layers with considerable differences in concentration of the several constituents of sea water has been observed by the Canadian scientists. On the other hand, the waters of San Juan Channel are relatively uniform. The erose nature of the channel's floor and the velocity of the tidal currents create very turbulent conditions.

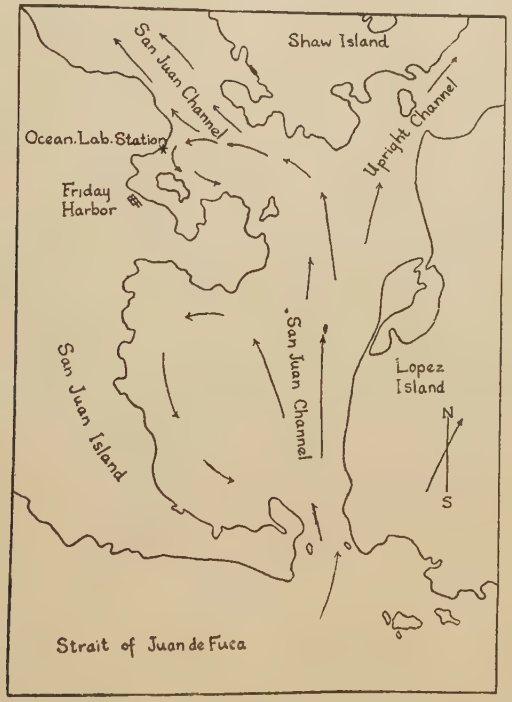

Figure 12. Chart of a part of San Juan channel showing the location of the Oceanographic Laboratories and general set of the currents during flood tides.

During the flood tide, the waters enter at the southern extremity of the Channel from Juan de Fuca Strait and at the northern extrance from Haro Strait. On the ebbing tide, conditions are reversed; Haro Strait and Rosario Strait are the passages through which interchange of water takes place between the Strait of Georgia and the Strait of Juan de Fuca. It is, therefore, only indirectly that the waters of the Strait of Georgia influence those of San Juan Channel, and then only for short periods of time.

METHODS OF OBSERVATION AND DETERMINATION

An observation pier of cantilever construction, projecting 27 meters from the shore over a maximum depth of 10 meters, was built to facilitate the taking of samples at regular intervals regardless of unfavorable weather conditions. A few meters beyond the end of the pier, the channel drops to a depth of about 130 meters. Fig. 13 is an illustration of the pier. Besides being equipped with the usual water sampling apparatus, the pier is equipped with a standard tide gauge, a constant recording thermometer for air temperature, automatic recorder for pyrheliometer readings, wet and dry bulb thermometers and a standard rain gauge near the entrance to the pier.

Figure 13. Cantilever Pier, Oceanographic Laboratories.

Surface samples taken for chemical analysis were secured at the peak of the higher flood tide, using a modified Ekman water bottle of four-liter capacity, equipped with Richter reversing thermometers. The schedule for sampling was every fourth day from October 1931 to June 1934 and every fifth day thereafter. Sampling has been fairly regular. Immediately after sampling, determinations of phosphate, nitrate, nitrite, and pH were made. The samples for oxygen, after the addition of sulfuric acid, and chlorinity were allowed to accumulate and these constituents determined at irregular intervals. The usual methods for the various determinations, as standardized by the Oceanographic Laboratories, were utilized except in the case of nitrates. In these determinations, the Harvey method was employed (*8*). In discussing this determination, Harvey (*9*) states: "The method provides good rough quantitative data, and since the differences are very large, such data are sufficient to give a picture of the distribution

of the nitrate in the sea." The authors are in accord with this statement, especially that characterizing the quantitative nature of the determination, as a large experimental error is very apparent.

TEMPERATURE

In Table I are given the mean monthly temperatures from January 1931 to December 1935. The seventh column of the table shows the mean of these mean temperatures for each particular month for the five-year period. The highest and the lowest temperatures recorded during this time for each individual month of the year are given in the eighth and ninth columns, respectively. Fig. 14 gives a graphic representation of the mean monthly temperatures during the period.

The lowest mean temperatures were obtained during the winter months. For February and March of both the years 1932 and 1933, the values were about one degree less than those observed for the other years. These lower temperatures were the result of prolonged subnormal air temperatures caused by an anticyclonic area centered to the north in western Canada. On February 17, 1932, the lowest temperature, 6.38° C., of the entire period occurred.

The highest mean temperatures were observed during July and August, 1933. It is interesting to note that these high values were preceded in June and followed in September by the lowest mean temperatures for these months during the five years. The highest temperature recorded, 14.25° C., was observed on August 15, 1933. These data indicate that for the months of July and August of this particular year, considerable water of much higher temperature was carried southward into San Juan Channel from the Georgia Strait.

The average yearly temperatures were remarkably constant and for three of the years did not vary more than 0.07° C. The highest yearly mean was that of 1931 with 9.23° C. and the lowest that of 1933 with 8.74° C.

The mean of the mean monthly temperatures for the five years is plotted in Fig. 15 together with the maximum and minimum values obtained during this period for each month of a year. The curves for the mean and minimum values are very regular. The same is true of the maxima except for July and August. This anomaly emphasizes again the effect of the diluted surface water of the Georgia Strait, which the Canadian investigators have shown to have summer temperatures about ten degrees higher than the surface waters of portions of the San Juan Archipelago, upon the waters of San Juan Channel.

CHLORINITY

In Table II are given the average monthly chlorinities of the surface waters from January 1931 to December 1935. The seventh column shows

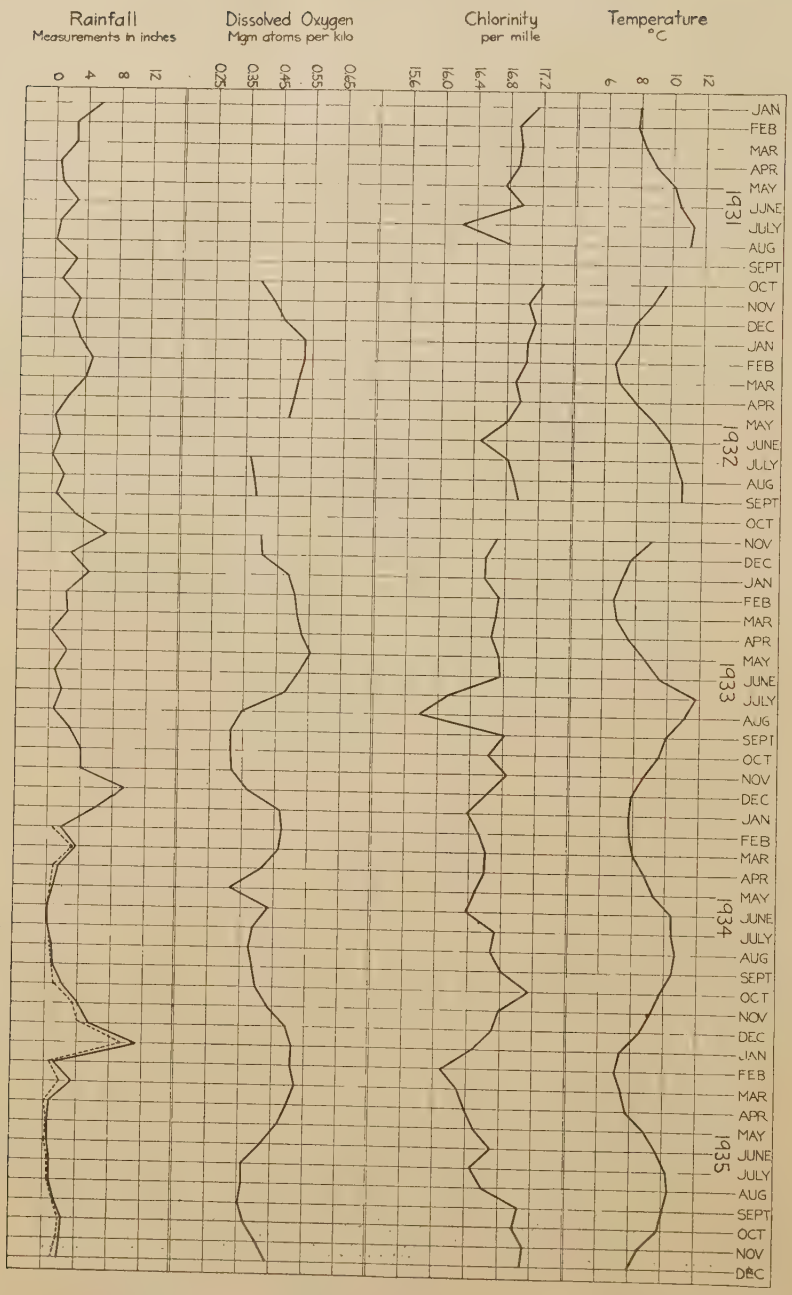

Figure 14. Graphs showing the mean monthly temperatures, chlorinities, dissolved oxygen values and rainfall data for the period, January 1931 to December 1935.

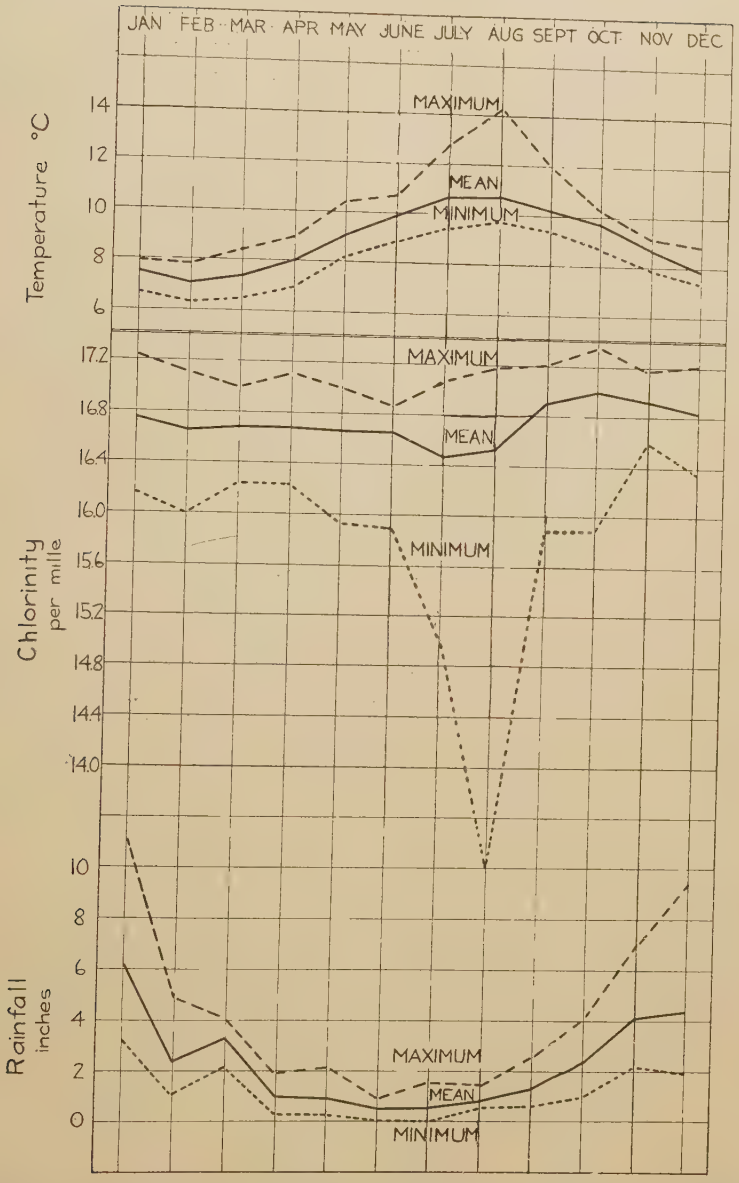

Figure 15. Graphs showing the mean of the mean monthly temperatures, chlorinities, and rainfall data for the five years, together with the maximum and minimum values obtained in each month during the period.

the mean of the mean chlorinities for each month of the year, while in the eighth and ninth columns are given the maximum and minimum values reported in each particular month for the entire period. The monthly mean values are plotted in Fig. 14. The effect of dilution by brackish water influenced by the discharge of the Fraser River, carried down from Georgia Strait in the summer months, is shown by the graph. It is interesting to note that this decrease in concentration occurs near the end of the period of minimum rainfall. The most pronounced dilution effect for the summer months for each year occurred for 1932 and 1934 in June, for 1931 and 1935 in July, and for 1933 in August. This effect was greatest in 1933 and the least in June, 1932. For a period of nearly two and a half years, that is, from March 1932 to September 1934, no monthly mean chlorinity values above 17°/$_{oo}$ were recorded, and most of the values during this period were below the average of all values for the five years. This may be attributed to the high rainfall experienced along the entire Oregon, Washington and British Columbia coasts; to heavy precipitations of snow in the coastal ranges with resulting increase in the volume of waters discharged by the rivers of the region; or to the effects of diluted waters from the Fraser River carried into San Juan Channel by prevailing northerly and northwest winds.

The graph also indicates the result of the winter rainfall, this effect being noted about January. The extent of the dilution effect in the winter months is a direct function of the rainfall for the entire region. Thus, in January 1935, abnormally heavy rains characterized the entire Puget Sound, Strait of Juan de Fuca, and British Columbia regions as well as the San Juan Archipelago. This abnormality is shown in Fig. 14. Along the coastal region, this precipitation was in the form of rain and resulted in immediate run-off with all rivers at flood stage. In the summer months the effect of the winter precipitation, which occurred in the form of snow in the mountains, has a marked influence upon the waters during the flood period of the river.

The highest chlorinities were always obtained in the fall of each year following the period of minimum rainfall. High chlorinities are generally accompanied by temperatures lower than average; low chlorinities are accompanied by temperatures higher than average. This general relationship is indicated in Fig. 15 which gives the graphs for the data given in the last three columns of Table II.

DENSITY

In Table III are given the average densities in situ for each month from January 1, 1931, to December 31, 1935. The densities follow very much the same trends as given previously for the chlorinities.

OXYGEN

In Tables IV and V are given the average values for dissolved oxygen for each month for the surface waters from October 1, 1931, to December 31, 1935. The values in Table IV are expressed as milligram atoms of oxygen per kilogram of water, while in Table V the values are given as per cent saturation. The seventh column in each table shows a mean of the mean dissolved oxygen values for each month of the year, while in the eighth and ninth columns are given maximum and minimum values obtained in each particular month for the entire period. The monthly mean values of Table IV are plotted in Figure 14. The data given in Table V would follow very much the same curve. The data shown in the seventh, eighth, and ninth columns of Table V are illustrated in Figure 16.

The highest oxygen values are obtained in the winter months and are in accord with other observations previously recorded from this region. These high oxygens are explained as being the result of strong wind and wave action during the winter months. Minute bubbles of air, contained in the water as the result of wave action, are carried below the surface as the result of the marked agitation by the tidal currents. With increasing pressure, the quantity of gas going into solution increases. Thus during the winter months, aeration from the atmosphere, in the waters of San Juan Channel, affects water layers at some depth below the surface. As the solubility of a gas increases with decreased temperature, another factor functions during the winter months which favors the rate of solution of greater quantities of oxygen. In Figure 14 it will be observed that the amplitude of the curves for dissolved oxygen varies from year to year, and this to a large measure may be attributed to weather conditions.

The winter and fall of 1935, the fall of 1933, and the winter of 1934 were characterized as being of considerable mildness, while during the fall months of 1934 and the winter months of 1935, there were strong winds of long duration accompanied at times by heavy rainfall. These conditions augmented by the turbulent nature of the waters had a marked influence upon the quantity of dissolved oxygen. The difference between these two periods is illustrated in Figure 14. With the coming of spring, there is generally a beginning of decreased oxygen values. The values then decrease generally throughout the summer and remain more or less constant during the fall months.

In the reports of previous investigations by the authors or their colleagues, the spring and fall maxima of oxygen values have been shown for the region of the San Juan Archipelago, Puget Sound, and the Strait of Juan de Fuca. Maxima were always obtained in the surface waters or waters near the surface. The data reported herewith are unique in that such maxima are not manifested. This apparent absence is explained by the fact that

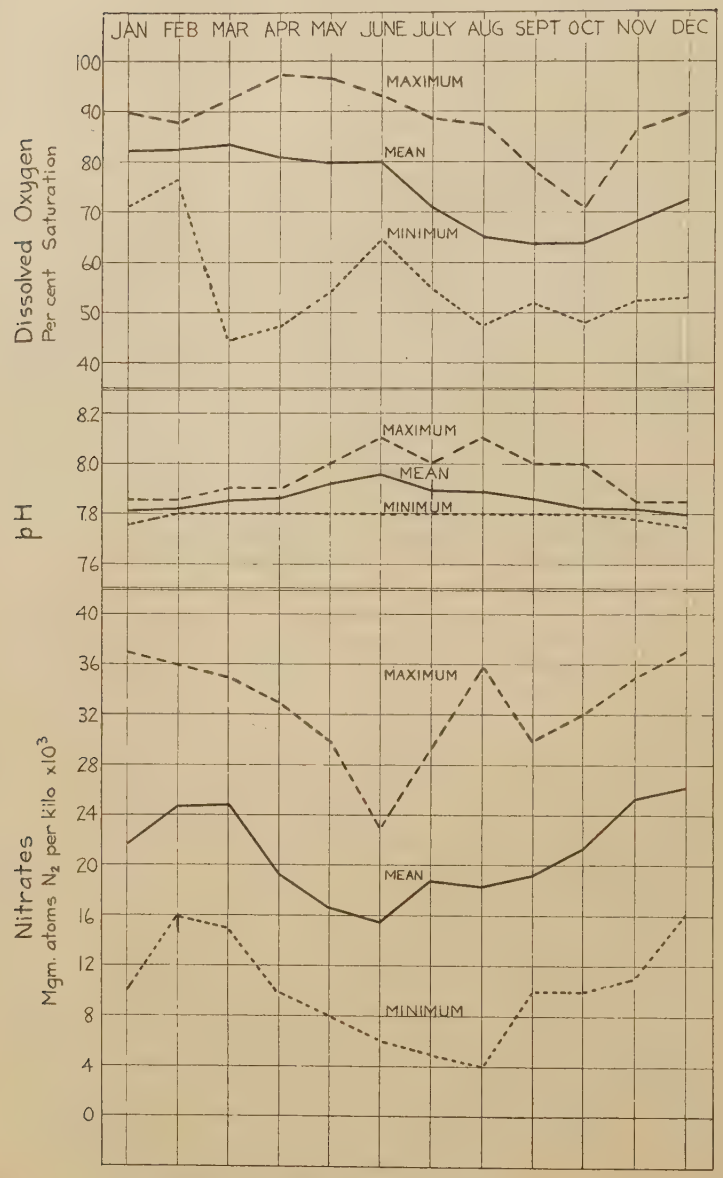

Figure 16. Graphs showing the mean of the mean monthly values for dissolved oxygen, pH, and nitrates, together with the maximum and minimum values obtained in each month during the five-year period.

samples were taken at a point where the waters had been subjected to the marked turbulence of tidal action in water layers averaging over a hundred meters in thickness. However, it should be stated that a plot of the detailed data would occasionally show high values for short periods of time in the spring and fall months. The plot of the maximum values in column eight of Table V, shown in Figure 16, indicates the effect of reactions due to photosynthetic activity more markedly than does the plot of the average values. It should be noted, however, that conditions approaching saturation are very seldom obtained.

The low oxygen values as indicated above may be explained by the fact that the first ten or twenty meters mark the zone where the process of photosynthesis takes place (*6*). In this zone, these reactions which tend to increase the dissolved oxygen, more than balance the reactions which tend to remove the dissolved oxygen. In the water layers below this depth, all conditions favor the removal of oxygen from solution. Thus, in the water layers of San Juan Channel, averaging 100 meters in depth, there are only about twenty meters of water favoring the increase of dissolved oxygen, while in the other eighty meters, factors exist which tend to lower the oxygen content. Considering conditions of turbulence and mixing, the results obtained are thus not surprising.

During the last three years of this series phytoplankton samples were taken jointly with those for oxygen. The phytoplankton maxima of production occurred during June 1933 and 1935, and July 1934. The correlation between the maxima of plankton and dissolved oxygen content is very small if any. In this period the dissolved oxygen content had been diminishing at least a month before the phytoplankton maxima occurred. This phenomenon is decidedly opposed to the usual case elsewhere, where the seasonal phytoplankton growth is accompanied by an increase in dissolved oxygen content, as cited by Dakin and Colfax (*3*), Harvey (*9*), Gran and Braarud (*5*), Braarud and Klem (*2*), Føyn (*4*).

During the period of rapid phytoplankton growth, the upper twenty meters of water tend to be enriched with oxygen, but general turbulence of the waters produced conditions that favor a mixing and a distribution of the oxygen throughout the entire water mass. Another factor is that the great abundance of zooplankton during the summer months in the upper water layer consumes much of the oxygen as the result of respiration; while, at the same time in the deeper water, the processes of respiration demand the consumption of quantities of oxygen very much greater than that required during the winter months. At the time of greatest biological oxygen production and consumption there is relatively little disturbance of the surface waters by wind and wave action and hence the processes of aeration experienced during the winter months are very considerably retarded. The increased temperature of the water over that of the colder portions of the year would tend to retard the rate of solution.

pH

The pH values are shown in Table VI and are characterized by their relative constancy throughout the period of observation. These values are not corrected for salt error, any such correction being practically a constant. Cresol red was the indicator used. The results are correct for individual determinations to at least $\pm$ 0.05 pH units.

Because of the small variations, the data for the period are not plotted. However, the monthly mean pH units are of interest when portrayed graphically as in Fig. 16. Constant values are shown for the winter months followed by increases in the spring and summer months, with decreases occurring in the fall. The graph for the maximum values shown in column eight follow very much the trend of the graph for the average values. On the other hand, the graph for the data in column nine is a straight line except for the low values of November, December, and January.

PHOSPHATES

In Table 7 are given the data showing the variations in the concentrations of phosphates. These data are plotted in Fig. 17.

There is a decided variation in the concentration of the phosphates from year to year. Such variations can not be attributed to differences in technique, to experimental error, or to possible changes in the standard solutions. The latter were always checked and prepared by at least two individuals, and the standard phosphate solutions used in the laboratories at Friday Harbor were always checked from time to time against the standard solutions prepared on board the research vessel M. S. CATALYST. The ship standards were prepared in turn by different individuals. Occasional checking of the methods used at the laboratory and aboard ship was performed on the same samples.

In the study of the seasonal variations of the phosphate concentration of the waters of San Juan Channel in 1929, (*23*), one of the authors obtained a mean yearly value of 0.0020 milligram atoms of phosphorous per kilo. This value is higher than most of the values obtained in the period from October 1931 to October 1935. From the data presented herewith, it must be concluded that the mean yearly phosphate concentration is by no means nearly as constant for other constituents and conditions reported above. The phosphate data tend to show a change from year to year, with the possibility of maximum values being obtained every four or five years.

In general it may be said, however, that the phosphates begin to decrease after February, the decrease becoming more marked during the spring and finally reaching the lowest values about June. The phosphate concentration of the waters during the summer months for 1933 remained fairly constant from June to October. In 1934, however, there was a decided

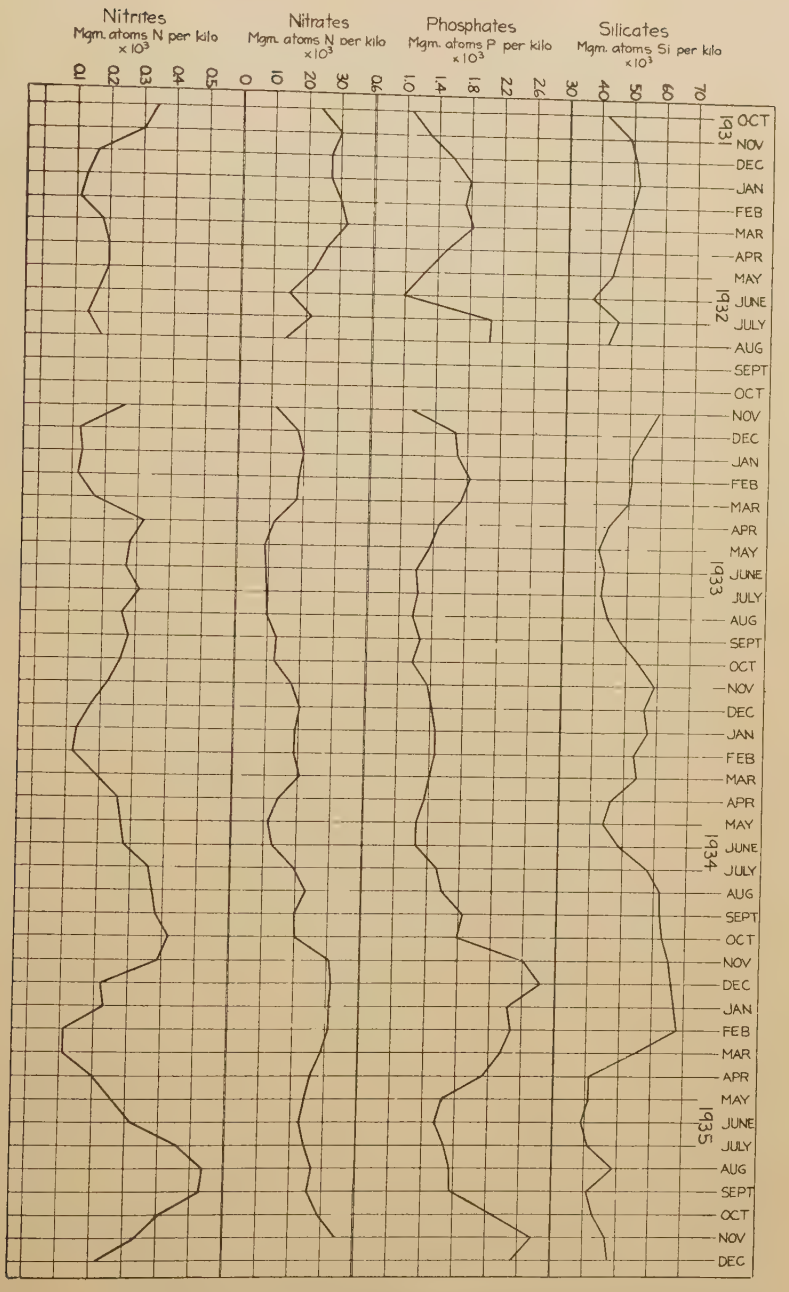

Figure 17. Graph showing the mean monthly concentrations for silicates, phosphates, nitrates and nitrites from October 1931 to December 1935.

tendency for increased values after the June minimum. In 1935 the concentration showed a tendency to increase after June but was not as marked as that observed during the previous year. Unusually high values were obtained in the fall of 1934 and the winter and fall of 1935. Graphs of the monthly means and the maxima and minima values as shown in the seventh, eighth, and ninth columns of Table 7 and illustrated in Fig. 18, indicate the general variation in the concentrations of the phosphates.

The seasonal decrease of the phosphate content of sea water due to the phytosynthetic activity of the phytoplankton followed by increased concentrations during the remainder of the year is common knowledge from diverse areas throughout the world. This gross relationship was found to coincide with the data of this investigation. The point of non-conformity with the general observation is simply that at no time did the phosphate content approach zero concentration, as has been noted in other localities. Harvey (*9*), and Gran and Braarud (*5*) emphasized the point that the phosphate content varies widely from one region to another and is dependent upon the "life history of the water." Moberg (*17*) and McEwen (*15*) have demonstrated the upwelling of bottom water along the North American coast. Thompson et al (*23*) expressed the opinion that the high phosphate content of this area is the result of upwelled water. The influx of this upwelled water into the Strait of Juan de Fuca, augmented by the normal enrichment as the result of leaching processes in the bottom waters of the area together with the markedly turbulent conditions, offers a plausible explanation of the fact that the phosphate concentration of the surface water was never depleted. In other words, the phosphate concentration was always sufficient for plant growth and thus can hardly be considered as one of the limiting factors.

SILICATES

Changes in the concentrations of the silicates during the five years are shown in Table VIII and the data graphically portrayed in Fig. 17. The values ranged from 40 to 60 mg. at. Si throughout the five years, except for the period from August 1934 to February 1935 when higher concentrations were observed. These latter occurred at practically the same time as the abnormally high phosphate values. The graph shows the seasonal fluctuations in the concentration of silicates. The data in the last three columns of Table VIII are plotted in Fig. 18 and show that silicates begin to decrease in concentration in February, the decrease continuing until June, after which there is a steady increase to November.

The graphs expressing the concentration of silicates follow the same general trend as those for phosphates. The ratio of milligram atoms of silicon to phosphorus averages 30, the deviation from the mean seldom being more than 10 per cent; however deviations as great as 50 per cent may

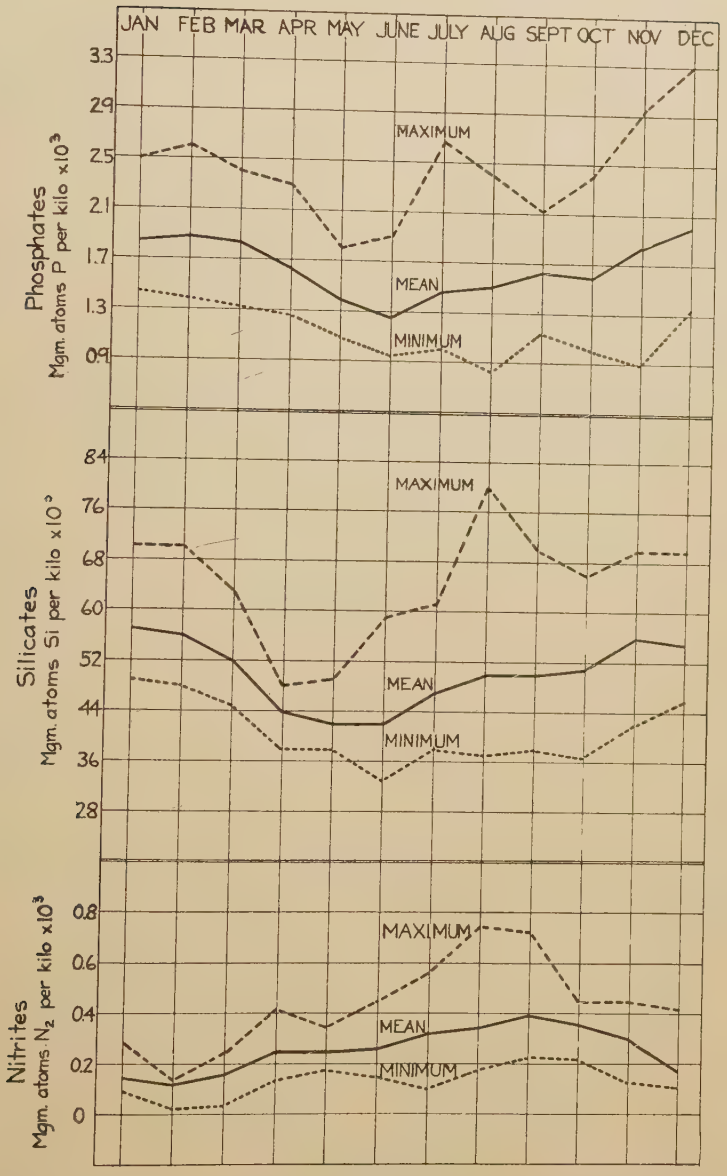

Figure 18. Graph showing the mean of the mean monthly values for phosphates, silicates, and nitrites, together with the maximum and minimum values obtained in each month during the five-year period.

sometimes be obtained in individual samples. Below are given the ratios calculated from the monthly mean values of Tables 7 and 8.

	Si/P		*Si/P*
January	.31	July	.32
February	.30	August	.33
March	.28	September	.31
April	.27	October	.32
May	.30	November	.31
June	.34	December	.28
		Mean	.30

NITRITES

Changes in the concentration of the nitrites in the five years are shown in Table IX and the data illustrated in Figs. 16 and 17. From a study of the graph in Fig. 17, it may be said that the concentration of nitrites tends to increase beginning with the summer months, the increase extending until early fall when the maximum is reached. After October, the concentration of nitrites rapidly decreases, finally reaching a minimum in February. This relationship is illustrated in Fig. 18 where curves for the mean maximum, and minimum average values for each month of the year are given.

There seems to be an inverse relationship between the concentration of nitrites and phosphates, which may be accounted for by the consumption of the phosphate by the phytoplankton and the correlation of the phytoplankton production maxima with the subsequent increased nitrite concentrations. The gross seasonal relationships between the phytoplankton activity and the nitrite content of this area are in general agreement with the work of T. Soot-Ryen (*20*) in the North-Norwegian fjords, N. Verjbinskaya (*25*) in Barents Sea and Atkins (*1*) in the English Channel. Also, the authors agree with the above writers that the varying nitrite content of sea water reflects an index of the oxidation of the nitrogen organic compounds to nitrate, which in turn re-enter the nitrogen cycle of the sea as nutrient salts usable by the phytoplankton population. The ratios of the mean milligram atom values for each month of the year of the concentration of phosphorus to the nitrogen of the nitrite show constancy about seven months of the year as illustrated below.

	N/P		*N/P*
January	.081	July	.21
February	.063	August	.23
March	.087	September	.25
April	.15	October	.23
May	.18	November	.16
June	.21	December	.09
		Mean	.15

NITRATES

The data for the nitrates are given in Table X and illustrated graphically in Figs. 16 and 17. Considerable difficulty is experienced in the determination of nitrates, using the reduced strychnine sulfate method. As stated previously, much of the data reported herewith has been carefully controlled and checked with standards prepared by various individuals, and in some cases, samples have been analyzed independently by different individuals. Excellent checks and agreements are invariably obtained for all the constituents studied with the exception of nitrates, and here the variations may sometimes be enormous.

Redfield (*18*) presents data showing the tendency for the existence of a constant ratio between the concentration of nitrates and phosphates. He states: "It is scarcely to be expected that these relations will hold in water bodies of limited extent, where local conditions such as the proximity of land, the inflow of rivers, and the peculiarities of a local flora and fauna may alter the picture. Examination of data for such regions as the Baltic, the Norwegian fjords, the Denmark strait, indicate that this is the case." According to data by Harvey (*9*), there is a variation in the English Channel of this ratio and Cooper has shown that in this region where active regeneration takes place, variation in the ratio of nitrate to phosphate is attributable to the longer time required to complete the regeneration of nitrate from decomposing albuminous matter. In the Gulf of Maine data recently secured by Dr. Rakestraw indicate that the ratio of the variation of nitrate and phosphate with depth is lower and more variable than in the open ocean outside the Gulf.

The data reported herein show a marked variation in the ratio for individual samples. However, the ratios of the average monthly values are numbers that vary less than ten per cent from the mean. These ratios are much smaller than those reported by Redfield.

	N/P		*N/P*
January	11.8	July	12.8
February	13.1	August	12.3
March	13.6	September	11.9
April	11.9	October	13.5
May	12.0	November	13.9
June	12.4	December	13.1
		Mean	12.7

Both the nitrate and phosphate values appear to be greater than those presented by investigators for various parts of the Atlantic. The concentrations of nitrate in San Juan Channel are not subjected to as marked a seasonal variation as manifested by other constituents. The authors feel that, due to the inadequacy of the determination of nitrates, it is unwise to

draw any definite conclusions at the present time. However, the data for the mean, maximum and minimum values for each month in the year demonstrate decided seasonal periodicity, with the tendency of the concentration of nitrates to decrease during the spring and early summer months and to increase thereafter, these variations being due to the photosynthetic activity of phytoplankton and to processes of oxidation.

In Table XI is given the precipitation received at the town of Friday Harbor at a point approximately one and a half miles from the Observation Pier where all other data was taken. These precipitation totals are expressed as inches of water and are plotted in Fig. 14. In the sixth column of the table is shown the mean precipitation for each month, while in the seventh and eighth columns are given the maximum and minimum values observed for each month of the entire period. All these are expressed graphically in Fig. 15. While heavier precipitation occurred during the winter months, the greatest value being 11.34 inches for January 1935, lighter rainfall was recorded during the summers, with July 1932 the only month showing no precipitation.

The rainfall in the San Juan Archipelago is somewhat less than that of the surrounding mainland. An explanation for this phenomenon may be that the prevailing southerly storms from the Pacific, which are responsible for much of the precipitation, lose most of their moisture in rising over the Olympic Mountains, and since the Archipelago lies in the lee of these mountains, it receives relatively less rainfall than contiguous territory.

Table XII gives a comparison, by monthly totals and differences, of the precipitation received for the years 1934 and 1935 at the town of Friday Harbor and at the Laboratories. It will be noted that the precipitation at the town exceeded that at the Laboratories with the exception of three months, June and July 1934 and October 1935. Nevertheless, the total precipitation for both years at the town was greater than that of the Laboratories by approximately 20 per cent. A recalibration of both gauges indicated these differences were actual. The occurrence of such differences within a distance of two miles and within 50 feet in elevation seemed interesting and is undoubtedly the result of local topography.

SUMMARY

1. Data are presented showing changes in temperature, chlorinity, and pH and changes in the concentrations of dissolved oxygen, phosphates, silicates, nitrites, and nitrates in the sea water near Friday Harbor, Washington, during a five-year period, January, 1931, to December, 1935.

2. The average yearly temperature of the surface waters for three of the five years did not vary more than .07° C. The highest yearly mean occurred in 1931, 9.23° C., and the lowest in 1933, 8.74° C.

3. The average chlorinity was 16.72‰. The highest chlorinities were obtained in the fall of the year and the lowest during the summer months.

4. The dissolved oxygen values are considerably different from those ordinarily reported from other regions. The surface waters seldom showed a saturation value greater than 80%. Local conditions are given as an explanation.

5. The pH values did not vary by more than .32 pH units throughout the entire period.

6. The phosphates showed a decided seasonal fluctuation, and the average yearly values indicate a periodicity extending over a number of years.

7. The silicate values likewise showed a seasonal variation and with the exception of the year 1934, when abnormally high values were obtained, the yearly averages are quite constant.

8. Highest nitrite values were obtained in the summer months and the lowest in the winter. Comparison of figures shows an inverse relationship exists between the nitrites and phosphates. The nitrite values show a decided seasonal variation as well as a variation from year to year.

9. Although the nutrient salts show an annual decrease during the period of greatest phytoplankton production, it is unique that their concentration did not diminish sufficiently to become a limiting factor in the phytoplankton productivity.

10. The effect of rainfall upon the sea water is discussed and data is presented showing the variation in precipitation between two stations separated by less than two miles.

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TABLES

TABLE I

THE AVERAGE TEMPERATURES FOR EACH MONTH OF THE SURFACE WATERS, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. JANUARY 1, 1931 TO DECEMBER 30, 1935.

	1931 °C	*1932* °C	*1933* °C	*1934* °C	*1935* °C	*Monthly Mean* °C	*Maximum* °C	*Minimum* °C
January	7.92	7.40	7.13	7.71	7.33	7.54	8.00	6.72
February	7.80	6.54	6.62	7.73	7.08	7.15	7.90	6.38
March	8.23	6.88	6.83	8.01	7.44	7.48	8.50	6.55
April	8.93	7.79	7.59	8.71	7.76	8.16	9.00	7.03
May	10.07	9.03	8.54	9.26	8.91	9.16	10.50	8.30
June	10.43	9.99	9.46	10.39	9.68	9.99	10.82	8.95
July	11.28	10.32	11.76	10.40	10.26	10.80	12.90	9.50
August	11.08	10.76	11.03	10.65	10.37	10.79	14.25	9.80
September		10.75	9.94	10.46	10.10	10.31	11.97	9.42
October	9.65		9.52	9.73	9.73	9.66	10.26	8.82
November	8.86	8.96	8.57	9.19	8.57	8.83	9.29	8.00
December	7.76	7.67	7.89	8.49	7.93	7.95	8.96	7.39
Yearly Mean	9.27	8.81	8.74	9.23	8.77	8.99		

Maximum of Maxima..... 14.25° August 15, 1933
Minimum of Minima..... 6.38° February 17, 1932
Difference..... 7.87°

TABLE II

THE AVERAGE CHLORINITIES FOR EACH MONTH OF THE SURFACE WATERS, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. JANUARY 1, 1931 TO DECEMBER 30, 1935.

	1931 °/₀₀	*1932* °/₀₀	*1933* °/₀₀	*1934* °/₀₀	*1935* °/₀₀	*Monthly Mean* °/₀₀	*Maximum* °/₀₀	*Minimum* °/₀₀
January	17.13	17.04	16.55	16.37	16.46	16.75	17.22	16.17
February	16.91	17.03	16.73	16.52	16.08	16.65	17.11	15.99
March	16.94	16.90	16.69	16.60	16.28	16.68	16.98	16.24
April	16.91	16.96	16.64	16.59	16.37	16.69	17.11	16.23
May	16.75	16.81	16.73	16.47	16.49	16.65	17.02	15.93
June	16.96	16.48	16.75	16.37	16.70	16.65	16.86	15.89
July	16.20	16.82	16.12	16.72	16.45	16.46	17.07	15.00
August	16.80	16.89	15.76	16.67	16.60	16.54	17.18	13.20
September		16.95	16.82	16.80	17.04	16.90	17.21	15.88
October	17.22		16.62	17.14	16.98	16.99	17.36	15.88
November	17.05	16.70	16.85	16.78	17.11	16.90	17.17	16.58
December	17.13	16.56	16.60	16.70	17.08	16.81	17.21	16.34
Yearly Mean	16.91	16.83	16.57	16.64	16.64	16.72		

Maximum of Maxima..... 17.36°/₀₀ October 30, 1931
Minimum of Minima..... 13.20°/₀₀ August 15, 1933
Difference..... 4.16°/₀₀

TABLE III

THE AVERAGE DENSITIES IN SITU FOR EACH MONTH OF THE SURFACE WATERS, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. JANUARY 1, 1931 TO DECEMBER 30, 1935.

	1931 σ_t	*1932* σ_t	*1933* σ_t	*1934* σ_t	*1935* σ_t	*Monthly Mean* σ_t	*Maximum* σ_t	*Minimum* σ_t
January	24.16	24.06	23.22	23.09	23.30	23.57	24.27	22.95
February	23.86	24.17	23.24	23.30	22.76	23.47	24.27	22.63
March	23.84	23.71	23.66	23.38	22.99	23.52	23.93	22.97
April	23.70	23.91	23.47	23.40	23.08	23.51	24.08	23.01
May	23.27	23.54	23.42	23.02	23.06	23.26	23.86	22.22
June	23.54	22.91	23.37	22.69	23.26	23.15	23.53	21.97
July	22.32	23.34	22.11	23.54	22.78	22.82	23.72	20.51
August	23.19	23.33	21.74	22.91	22.80	22.79	23.77	17.58
September		23.43	23.40	22.93	23.88	23.41	23.98	21.52
October	24.00		23.18	23.24	23.62	23.51	24.27	22.08
November	23.91	23.38	23.64	22.64	23.98	23.51	24.15	22.50
December	24.15	23.36	23.41	22.76	24.06	23.55	24.23	22.42
Yearly Mean	23.72	23.56	23.16	23.08	23.30	23.51		

Maximum of Maxima.....24.27 January 17 and October 30, 1931 and February 25, 1932
Minimum of Minima.....17.58 August 15, 1933
Difference..... 6.69

TABLE IV

THE AVERAGE VALUES FOR DISSOLVED OXYGEN OF THE SURFACE WATERS FOR EACH MONTH, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1935.

Milligram Atoms per Kilogram of Water

	1931	*1932*	*1933*	*1934*	*1935*	*Monthly Mean*	*Maximum*	*Minimum*
January		0.526	0.485	0.466	0.513	0.498	0.542	0.428
February		.525	.505	.474	.510	.504	.540	.464
March		.518	.512	.465	.522	.504	.550	.250
April		.495	.526	.414	.498	.483	.581	.276
May		.480	.554	.370	.470	.469	.568	.314
June			.520	.434	.421	.458	.539	.366
July		.364	.477	.389	.362	.398	.506	.312
August		.375	.348	.376	.363	.366	.479	.269
September		.383	.314	.388	.354	.360	.432	.288
October	0.387		.314	.398	.371	.368	.416	.274
November	.430	.398	.318	.435	.409	.398	.502	.308
December	.462	.453	.366	.492	.441	.443	.544	.314
Yearly Mean		.451	.437	.425	.436	.437		

Maximum of Maxima.....0.581 April 18, 1933
Minimum of Minima..... .250 March 3, 1934
Difference..... .331

TABLE V

THE AVERAGE VALUES FOR PER CENT SATURATIONS OF DISSOLVED OXYGEN OF THE SURFACE WATERS FOR EACH MONTH, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1935.

	1931 %	*1932* %	*1933* %	*1934* %	*1935* %	*Monthly Mean* %	*Maximum* %	*Minimum* %
January		87.5	79.4	77.3	84.7	82.2	89.6	71.0
February		85.3	82.1	78.7	83.1	82.3	87.7	76.2
March		84.9	83.7	78.0	86.0	83.2	92.0	41.9
April		82.9	87.5	70.3	82.6	80.8	97.0	47.0
May		81.8	93.8	61.6	80.8	79.5	96.3	53.8
June			90.1	76.3	73.2	79.9	93.0	64.4
July		64.5	85.9	68.6	63.9	70.7	88.4	54.7
August		67.0	61.7	66.7	64.1	64.9	87.1	47.3
September		68.4	54.9	69.1	62.0	63.6	78.0	51.7
October	66.2		54.5	69.5	64.4	63.7	70.4	48.1
November	73.9	68.3	54.1	74.9	69.1	68.1	86.1	52.4
December	77.5	75.5	61.1	82.8	73.7	74.1	89.6	53.0
Yearly Mean		76.6	74.0	72.8	74.0	74.4		

Maximum of Maxima.....97.0% April 18, 1933
Minimum of Minima.....41.9% March 3, 1934
Difference.....55.1%

TABLE VI

THE AVERAGE pH OF THE SURFACE WATERS FOR EACH MONTH, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1935.

	1931	*1932*	*1933*	*1934*	*1935*	*Monthly Mean*	*Maximum*	*Minimum*
January		7.83	7.79	7.80	7.80	7.81	7.85	7.75
February		7.85	7.84	7.80	7.80	7.82	7.85	7.80
March		7.85	7.85	7.82	7.87	7.85	7.90	7.80
April		7.85	7.87	7.83	7.90	7.86	7.90	7.80
May		7.90	7.93	7.87	7.96	7.92	8.00	7.80
June		8.00	7.96	7.92	7.97	7.96	8.10	7.80
July		7.90	7.93	7.85	7.86	7.89	8.00	7.80
August		7.92	7.90	7.88	7.84	7.89	8.10	7.80
September		7.92	7.82	7.83	7.85	7.86	8.00	7.80
October	7.85		7.82	7.82	7.80	7.82	8.00	7.80
November	7.82	7.85	7.80	7.81	7.80	7.82	7.85	7.78
December	7.80	7.80	7.80	7.80	7.80	7.80	7.85	7.75
Yearly Mean		7.88	7.86	7.84	7.85	7.86		

Maximum of Maxima.....8.10 August 15, 1933
Minimum of Minima.....7.78 December 27, 1932
Difference..... .32

TABLE VII

THE AVERAGE CONCENTRATION OF PHOSPHATE OF THE SURFACE WATERS FOR EACH MONTH, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1935.

Milligram Atoms $P \times 10^3$ per Kilogram of Water

	1931	*1932*	*1933*	*1934*	*1935*	*Monthly Mean*	*Maximum*	*Minimum*
January		1.79	1.68	1.47	2.42	1.84	2.50	1.44
February		1.73	1.84	1.47	2.47	1.88	2.60	1.38
March		1.83	1.73	1.41	2.35	1.83	2.40	1.32
April		1.51	1.47	1.35	2.14	1.62	2.30	1.26
May		1.25	1.36	1.26	1.64	1.38	1.80	1.08
June		0.99	1.20	1.26	1.56	1.25	1.90	.94
July			1.22	1.52	1.68	1.47	2.66	1.00
August			1.17	1.59	1.75	1.50	2.40	.82
September			1.26	1.85	1.76	1.62	2.11	1.13
October	1.07		1.17	1.79	2.28	1.58	2.40	1.00
November	1.29	1.10	1.36	2.61	2.77	1.83	2.94	.89
December	1.58	1.65	1.42	2.82	2.53	2.00	3.29	1.33
Yearly Mean		1.48	1.41	1.70	2.11	1.65		

Maximum of Maxima.....3.29 December 3, 1934
Minimum of Minima.....0.82 August 11, 1933
Difference.....2.47

TABLE VIII

THE AVERAGE CONCENTRATION OF SILICATES OF THE SURFACE WATERS FOR EACH MONTH OF THE YEAR, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1935.

Milligram Atoms $Si \times 10^3$ per Kilogram of Water

	1931	*1932*	*1933*	*1934*	*1935*	*Monthly Mean*	*Maximum*	*Minimum*
January		52	51	57	67	57	70	49
February		50	51	53	68	56	70	48
March		48	50	54	55	52	63	45
April		46	44	46	41	44	48	38
May		44	41	44	41	42	49	38
June		38	43	49	39	42	59	33
July		46	42	58	41	47	61	38
August		43	44	62	49	50	80	37
September			48	62	41	50	70	38
October	42		54	63	43	51	66	37
November	49	59	59	65	47	56	70	42
December	51	55	56	66	48	55	70	46
Yearly Mean		48	49	57	48	50		

Maximum of Maxima.....80 August 30, 1934
Minimum of Minima.....33 June 15, 1932
Difference.....47

TABLE IX

THE AVERAGE CONCENTRATION OF NITRITES OF THE SURFACE WATERS FOR EACH MONTH, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1933.

Milligram Atoms of Nitrogen $\times 10^3$

	1931	*1932*	*1933*	*1934*	*1935*	*Monthly Mean*	*Maximum*	*Minimum*
January		0.13	0.13	0.13	0.23	0.15	0.29	0.10
February		.11	.12	.12	.11	.12	.14	.03
March		.18	.17	.19	.11	.16	.24	.04
April		.20	.32	.26	.20	.25	.42	.14
May		.20	.28	.27	.26	.25	.35	.18
June		.17	.27	.28	.32	.26	.45	.15
July		.14	.31	.36	.46	.32	.56	.11
August		.18	.26	.37	.54	.34	.75	.18
September			.28	.38	.53	.40	.73	.23
October	0.34		.26	.42	.41	.36	.45	.22
November	.30	.26	.22	.39	.34	.30	.45	.13
December	.16	.12	.17	.22	.22	.18	.42	.11
Yearly Mean		.17	.23	.28	.31	.26		

Maximum of Maxima.....0.75 August 29, 1935
Minimum of Minima..... .03 February 9 and 13, 1932
Difference..... .72

TABLE X

THE AVERAGE CONCENTRATION OF NITRATES OF THE SURFACE WATERS FOR EACH MONTH, OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES, FRIDAY HARBOR, WASHINGTON. OCTOBER 1, 1931 TO DECEMBER 30, 1935.

Milligram Atoms of Nitrogen $\times 10^3$

	1931	*1932*	*1933*	*1934*	*1935*	*Monthly Mean*	*Maximum*	*Minimum*
January		27.0	20.0	19.0	31.0	21.8	37	10
February		30.1	18.6	18.9	31.0	24.7	36	16
March		32.0	18.1	20.7	28.7	24.9	35	15
April		26.0	11.4	14.1	25.6	19.3	33	10
May		22.2	8.8	11.4	24.0	16.6	30	8
June		17.3	9.3	12.7	22.6	15.5	23	6
July		22.0	9.6	19.5	24.2	18.8	29	5
August		14.0	9.7	23.1	26.6	18.4	36	4
September			12.7	19.8	25.5	19.3	30	10
October	24.0		12.2	20.2	28.8	21.3	32	10
November	30.0	13.9	17.7	30.9	34.4	25.4	35	11
December	27.0	18.2	20.4	31.5		26.3	37	16
Yearly Mean		22.2	14.0	20.2	27.5	21.0		

Maximum of Maxima.....37 December 30, 1931 and January 3, 1932
Minimum of Minima..... 4 August 3, 1932
Difference.....33

TABLE XI

RAINFALL AT FRIDAY HARBOR FROM OCTOBER 1, 1931 TO DECEMBER 31, 1935

Measurements in Inches

	1931	*1932*	*1933*	*1934*	*1935*	*Average*	*Maximum*	*Minimum*
January	3.64	3.29	4.78	5.94	11.34	5.80	11.34	3.29
February	2.55	4.87	1.95	1.62	1.06	2.41	4.87	1.06
March	2.61	4.06	2.14	3.48	3.40	3.14	4.06	2.14
April	0.57	1.89	0.26	1.27	0.57	0.91	1.89	0.26
May	0.83	0.27	2.15	0.71	0.37	0.87	2.15	0.27
June	2.71	0.92	0.66	0.04	0.43	0.95	0.92	0.04
July	0.65	0.00	1.57	0.14	0.74	0.62	1.57	0.00
August	0.13	1.47	0.59	0.69	0.74	0.72	1.47	0.59
September	2.71	0.60	2.64	0.85	1.32	1.62	2.64	0.60
October	0.99	2.92	4.00	2.08	2.39	2.48	4.00	0.99
November	3.29	6.87	3.96	4.17	2.16	4.09	6.87	2.16
December	2.30	2.49	9.43	5.57	1.93	4.34	9.43	1.93
Total	22.98	29.65	34.13	26.56	26.45	27.95		

TABLE XII

COMPARISON OF THE RAINFALL IN THE TOWN OF FRIDAY HARBOR AND AT THE OBSERVATION PIER, OCEANOGRAPHIC LABORATORIES.

Inches of Rainfall

	1934			*1935*		
	Town	*Laboratories*	*Difference*	*Town*	*Laboratories*	*Difference*
January	5.94	Incomplete		11.34	9.47	−1.87
February	1.62	0.81	−0.81	1.06	0.72	−0.34
March	3.48	3.24	−0.24	3.40	1.98	−0.42
April	1.27	0.84	−0.43	0.57	0.16	−0.41
May	0.71	0.48	−0.23	0.37	0.31	−0.06
June	0.04	0.07	+0.03	0.43	0.21	−0.22
July	0.14	0.16	+0.02	0.74	0.71	−0.03
August	0.69	0.55	−0.14	0.74	0.54	−0.20
September	0.85	0.78	−0.07	1.32	1.81	+0.49
October	2.08	1.10	−0.98	2.39	2.04	−0.35
November	4.17	3.43	−0.74	2.16	1.84	−0.32
December	5.57	4.05	−1.52	1.93	1.08	−0.85
Totals	26.56	21.45*		26.45	20.87	

* The town January rainfall was substituted for the Laboratories for January.

THE SIGNIFICANCE OF THE MISSISSIPPI RIVER DRAINAGE FOR BIOLOGICAL CONDITIONS IN THE NORTHERN GULF OF MEXICO

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Plankton collections and bottom hauls made during the 1935 expedition of the "Atlantis" to the Gulf of Mexico suggested that there is a zone of high productivity in the shallow water area near the mouth of the Mississippi, and since the effect of river outflow on ocean life is a phase of marine biology which has not yet received adequate attention, further investigation seemed desirable.

Atkins (1926) published tables of the phosphate distribution in the vicinity of Plymouth Sound which show that in that region the rivers contribute little phosphate to the ocean, and in the same year Harvey obtained similar results for nitrate. But Atkins had pointed out before (1923) that in certain regions the incoming nutrients may be highly significant. His conclusion was based largely on the data of McHargue and Peter (1921) who estimated that in the Mississippi River the amount of soluble phosphorus liberated each year is about 62,188 tons, and, in addition, the suspended matter, which averages 7469 million cubic feet per annum, contains .15% P in combined form.

Harvey (1928) again pointed out that in the rivers of the Plymouth Sound region the nutrients are largely exhausted by plant growth in the tidal or even the non-tidal zone, but he suggested that it was probably not the case in a large river such as the Amazon. Hesse, Allee, and Schmidt (1937) concluded that coastal effluents are quite important and even suggested that they are in part responsible for the fact that the Atlantic, which receives a large proportion of the earth's drainage, has a higher nutrient content than the Pacific.

The present work was therefore undertaken during the 1937 cruise of the "Atlantis," jointly sponsored by the Woods Hole Oceanographic Institute and Yale University (Bingham Oceanographic Laboratory), in an attempt to determine whether or not the incoming nutrients from the Mississippi have an observable quantitative effect on plant growth and nutrient matter present in the Gulf of Mexico.

I want to express my gratitude to Dr. Albert E. Parr for giving me the opportunity to work on this problem and for his many valuable suggestions. I am also indebted to Mr. A. A. Hirsch of the New Orleans Sewerage and Water Board Company for much information and advice about the Mississippi River.

METHODS

The water samples were collected with Nansen bottles in the usual way, except stations B to H, which were surface samples drawn with a bucket thrown from the rail of the ship. Preparation of plankton collections and analyses of chlorophyll, soluble P, and soluble copper were done immediately; plankton weights and the remainder of the chemical analyses were obtained later from preserved samples. The positions of the stations where data of this sort were collected are shown in figure 19.

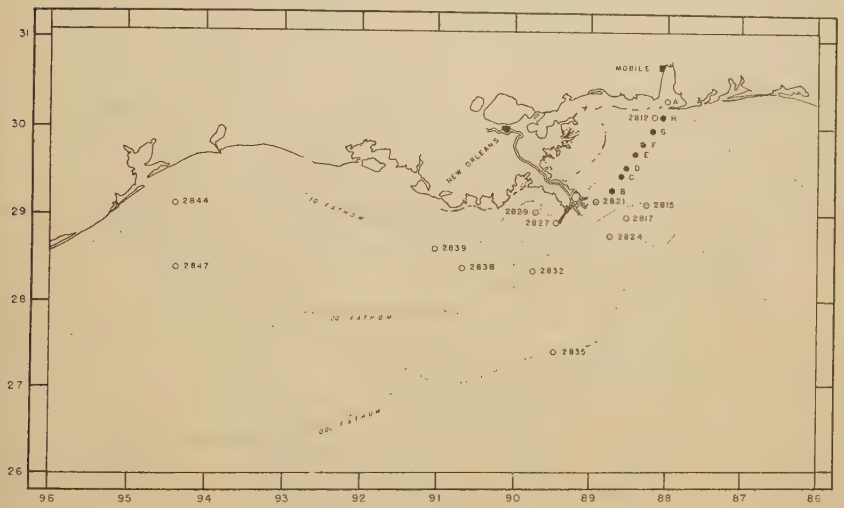

Figure 19. Position of stations referred to in this article.

Phosphate analyses were made according to the procedure described by Atkins (1923).

The copper test is a colorimetric determination described by Callan and Henderson (1929). A solution of sodium diethyl-dithiocarbamate is added to the sample to be tested. The copper salt of this substance is extracted with amyl alcohol, as suggested by McFarlane (1933), and for the colorimetric comparison I use a .001 N solution of potassium dichromate, which has the same color, the depth of which varies in a linear manner with that of the copper solution. When 50 ml. of water are used for the test plus .5 ml. of the reagent and 4.97 ml. of amyl alcohol, the .001 N. $K_2Cr_2O_7$ is the colorimetric equivalent of a copper concentration in the sample of .030 mg. per liter. The procedure of Haddock and Evers (1932) for the removal of iron is optional. Iron contamination is readily detected because it has a slightly reddish tint. It is unstable in amyl alcohol, fading after four or

five hours, whereas the copper color persists several days, so that it is possible to determine the copper content after the iron contamination has disappeared.

The chlorophyll was determined by a modification of Harvey's method (1934). The phytoplankton was removed from a liter or more of water by means of a quantitative filter paper and after separation of the chlorophyll in saponified form (Riley, 1937), it was measured colorimetrically against the Harvey standard.

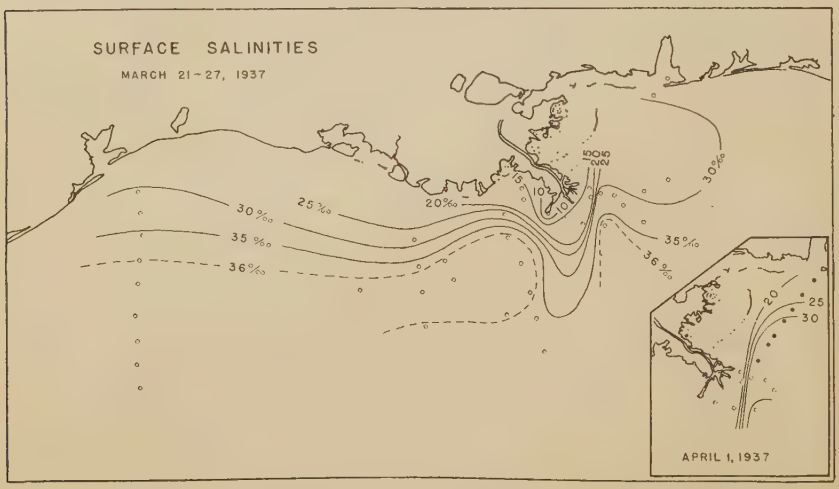

Figure 20. Distribution of surface salinities.

Plankton collections were made for the purpose of determining the weight of organic matter, and in some cases for copper and phosphate tests. One or two liters of water were filtered through a thirty-five second membrane filter, which removes all particulate matter of a diameter greater than .6 micron. The material was scraped from the filter after having been washed with distilled water to remove as much of the salt as possible, and was preserved with formalin. It was afterwards washed again, dried for two two days at 60° C., weighed, ignited, and weighed again, the difference being approximately the weight of organic matter in the sample. Copper tests were made on the ash. In the cases where phosphate tests were made, it was necessary to oxidize the whole sample, since ashing destroys part of the phosphate.

PHOSPHATES

The values obtained for soluble phosphorus in the surface waters varied from 2 to 16 mg. per m^3. The latter values were obtained at the mouth of

the Mississippi near Southwest Pass and Pass à Loutre, and the next highest figure, 11 mg. per m³., was in Mobile Bay. In the Gulf itself, the analyses ranged from 2 to 8 mg., with an average of 3.8. Figure 21 shows the general distribution of phosphates in the surface layer. The higher values are found around the mouth of the Mississippi and extending eastward toward Mobile Bay and westward to about longitude 91°.

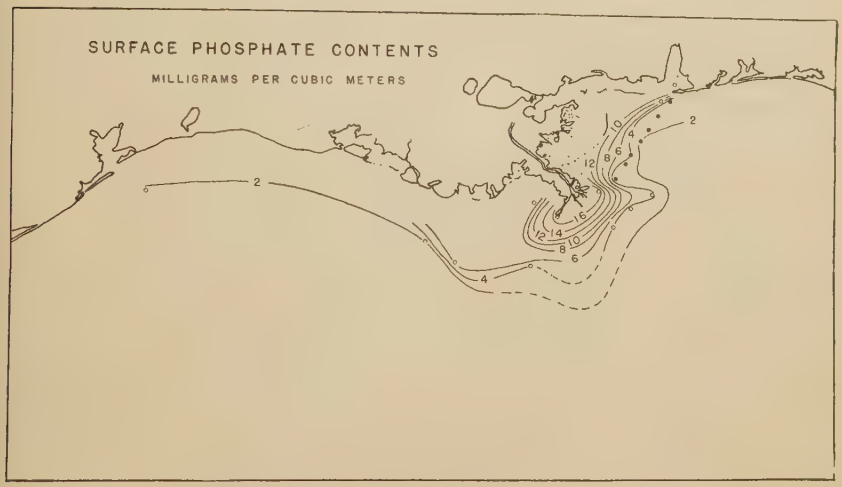

Figure 21. Phosphate distribution in surface layer.

This zone of high phosphate concentration is quite superficial. There are five stations within the zone where analyses were made at depths ranging from 10 to 25 meters. At those depths the mean amount of soluble P was 3.6 mg. per m³. as compared with 5.8 at the surface at the same stations. A similar result was obtained at Pass à Loutre, where the phosphate was 16 mg. at the surface and 11 mg. at 13 m. In Mobile Bay, on the other hand, it was 11 mg. at the surface and 15 at 6 m. Outside the zone of high surface phosphates there are only two stations available for study of vertical distribution, and their results are contradictory, so that no generalizations can be made.

PLANKTON MEASUREMENTS

The chlorophyll data are expressed in the arbitrary Harvey units, of which one unit corresponds to about .88 gamma of chlorophyll. It is impossible to convert these figures into absolute weights of phytoplankton, but the chlorophyll analysis serves as an index of potential photo-synthetic activity. As such, it is more useful in studies of productivity than plankton

weights, which are of course measurements of animals, bacteria, and detritus, as well as phytoplankton.

The mean amount of chlorophyll in all analyses of Gulf water was 6.7 Harvey units per liter. In this case also, it is possible to draw a rough zonation around the mouth of the Mississsippi. The region where the chlorophyll is highest, above 8 units per liter, extends westward from Mobile Bay to some undetermined point beyond station 2839. If it goes as far as Galveston, it is reduced to a narrow strip near the shore. The areas containing 6 to 8 Harvey units per liter form thin bands around the eight

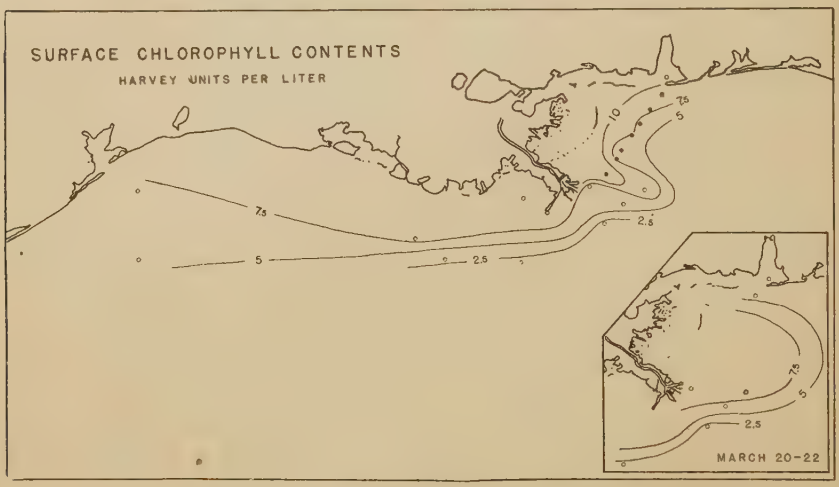

Figure 22. Chlorophyll distribution in surface layer.

unit line and extend westward past its observed limits. There remain four stations on the east and south, where the range of chlorophyll was from 1.6 to 5.9.

Station 2812 does not fit in with the remainder of the analyses. A surface value of 3.6 Harvey units per liter was obtained there, whereas in the waters immediately around it on the east, north, and south, the chlorophyll measured eight or more units. This, however, can be ignored because of evidence based on salinity distribution. According to Dr. Parr, the surface salinities show two distinct situations, observations during the early part of the cruise giving a consistent picture, which, however, cannot be combined with observations made on the return trip between the Mississippi and Mobile (see fig. 20). Heavy weather intervening between these two parts of the cruise probably caused the change. Since most of the special observations in this area were made on the return trip, most of the data on chloro-

phyll, phosphate, and copper distributions between Mobile and the mouth of the Mississippi river correspond to the salinity distribution of April 1. Only in the case of chlorophyll does it seem necessary to give two separate interpretations.

The only sample of completely fresh water, which was taken at New Orleans, had a value of 12.8 units per liter. This was slightly higher than any of the analyses made in the Gulf but was by no means large by comparison with the usual values obtained in fresh water. It is obvious that the turbidity of the Mississippi is a strong inhibiting factor on the growth of phytoplankton, which would otherwise be very great because of the large quantities of available nutrients. The five analyses near the mouth of the Mississippi and in Mobile Bay ranged from 4.8 to 11.1. Here also, turbidity was an important conditioning factor, and, in addition, the salinities were probably unfavorable.

The results obtained from the plankton weights closely parallel the chlorophyll measurements. The dry weights of organic matter per liter of water varied from 1.0 to 20.6 mg. in Mobile Bay and near the mouth of the Mississippi. In the Gulf the range was from .2 to .5 at the surface, the lowest values of .2 and .3 mg. being at stations 2832 and 2838, where the chlorophyll was also very low.

The very high ratios of organic matter to chlorophyll which occur near the river indicate large quantities of organic detritus. The ratios fall so rapidly as one proceeds out into the Gulf that it seems likely that practically all the organic detritus of fresh water origin is removed from the surface water before it gets more than ten or fifteen miles from the mouth of the river.

The plankton was not identified carefully, but superficial examination indicated that the flora is fairly constant qualitatively throughout the entire area studied in spite of the wide range of salinities. No freshwater phytoplankton was noted in the water at station 2827, which had a salinity of only 5.95‰.

COPPER

The work on copper was undertaken because it was thought that it might supply evidence on the nutrient conditions in the Gulf. In an as yet unpublished investigation of the copper cycle in fresh water it was shown that copper is readily absorbed on organic detritus, and the amount so combined is frequently much greater than that occurring in the living organisms of the plankton. It can therefore be used as an approximate index of the amount of detritus in a given sample of seston.* But since it

* A general term, borrowed from freshwater ecology, for the combination of plankton and detritus which is removed from the water by the usual filtration and centrifuge methods.

is apparent that freshwater detritus is not a very important source of nutrients in Gulf waters, the work on copper is not significant from that standpoint.

There are, however, certain aspects of the copper problem that are in themselves worthy of brief mention. The presence of copper in sea water has been detected by Dieulafait, Rose and Bodansky, Prytherch, and others. Prytherch found that the quantity in the sea is very much less than that in fresh water and suggested that inorganic precipitation takes place. The horizontal distribution of copper in the Gulf, shown in figure 23, confirms

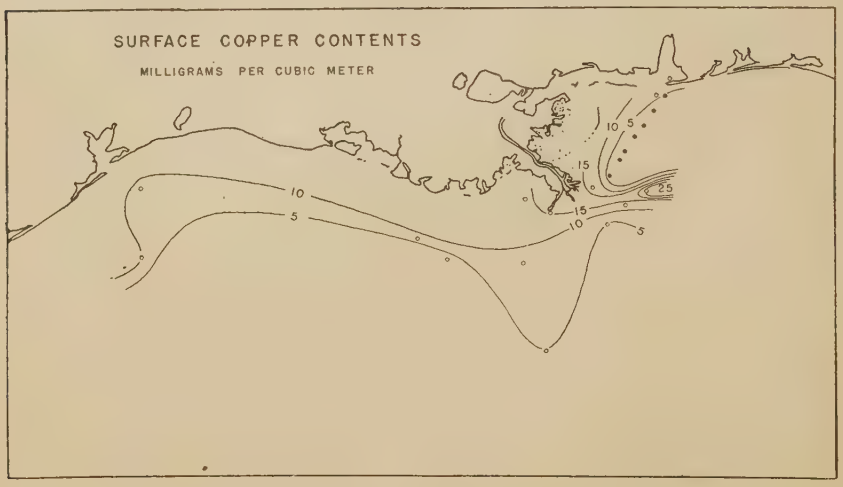

Figure 23. Distribution of copper in surface layer.

his results in so far as the amount near the Mississippi is large, and it decreases progressively with increasing salinity.

The total range of soluble copper in all analyses was 1 to 25 mg. per m^3. The very high values of 20–25 mg. obtained at station 2815 do not fit with the rest of the picture. It is not possible to say definitely whether or not there was an error in these analyses, but too much significance should not be attached to a single set of isolated values. Except for this station, the range of copper was from 1 to 15 mg., with the highest values in water of low salinity. Furthermore, analyses of the vertical distribution of soluble copper near the mouth of the river show that the amount is greater in the low salinity surface layer than at the lower levels where the salinity is higher. But none of the analyses obtained in the Gulf were negative, indicating that removal by precipitation, if it occurs at all, is not complete. It thus seems probable that copper is not a limiting factor in the open ocean to those animals which require it for the synthesis of hemocyanin.

At station 2835 tests were made to determine the vertical distribution of soluble copper. A slight accumulation was noted in the hypothalassa (fig. 24), which was much less than that recorded for such substances as phosphate and nitrate, which are biologically more active. The data on

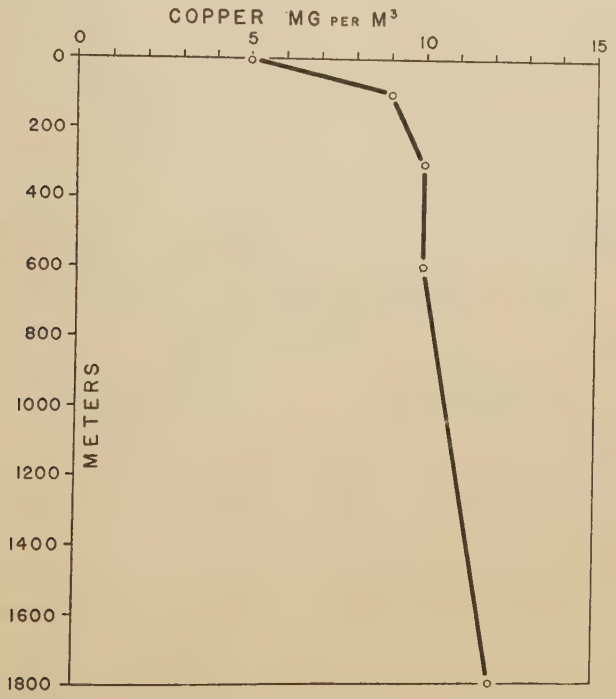

Figure 24. Vertical distribution of copper at station 2835.

vertical distribution and seston copper both resemble conditions found in freshwater lakes and indicate that the distribution factors are similar, with the emphasis placed on adsorption equilibria between soluble and seston copper, rather than on inorganic precipitation.

DISCUSSION

The nutrient content of the Mississippi River

Analyses of soluble phosphorus at stations 2821 and 2827 near the mouth of the river where mixing with sea water was slight gave values of 16 mg. per m^3. at the surface. At the former station 11 mg. were found at a depth of 13 m. One analysis was made of soluble P in the river proper, where a

value of 6 mg. per m^3. was obtained at Carrollton, just above New Orleans, on April 5. This difference cannot reasonably be accounted for on the basis of regeneration of combined phosphorus or discharge from the New Orleans sewerage disposal plant. The former is probably not more than 2 mg. per m^3.; the latter, according to an estimate by Mr. Hirsch based on the normal amount of phosphorus in human excreta and on the amount of phosphate washing powders used in New Orleans, is about 1500 kilos per day, which at the rate of flow at that time would average only .6 mg. per m^3. The difference is probably not important, for, as Mr. Hirsch pointed out, the nitrate content of the water above New Orleans varies irregularly, and it is quite possible that phosphate behaves in a similar manner.

The amount of phosphorus in combined form appears to be less than half as much as values obtained for soluble P at the mouth of the river. Analyses of sediments taken from the river at Carrollton are shown in the table below:

	Mg. of P/g. of mud
June 20–July 11, 1936	.011
October 10–30, 1936	.004
Dec. 12, 1936–Jan. 9, 1937	.053
January 9–31, 1937	.0008
Jan. 31–Feb. 6, 1937	.010
Feb. 6–Mar. 13, 1937	.006
Mar. 13–27, 1937	.007
Mar. 27–May 1, 1937	.004

These sediments are Berkefeld residues of weekly composite water samples collected twice daily and thus represent a mass average for a given period. The average of all samples is .012 mg. of P per gram of mud, and since the amount of sediment is usually less than a gram per liter (the most recent average that I have is .65 g./l. in 1935), it appears that the amount of combined phosphorus is rather small. At station 2827 it was .005 mg. per gram of mud, which was .002 mg. less than the amount at New Orleans during the same period. The weight of sediment collected at that station was 294 mg. per liter, so that the amount of combined P was .0016 mg. per liter, only one-tenth the value of the soluble P.

The discharge of the river at that time was 34,000 m^3. per second. If the data obtained at station 2827 may be regarded as accurate enough to be used as an estimate, the Gulf was at that time receiving 47,000 kilos of phosphate in soluble form per day, and an additional 4,700 kilos of combined P.* The latter figure is probably too low, for some of the mud had already become sedimented, but the value would almost certainly not be more than 25,000.

* E. g. enough to contribute one mg. per m^3. daily to 1000 square kilometers of surface layer, 50 meters deep.

It is obvious that these values represent only rough approximations. The averages obtained by McHargue and Peter in 1921 for soluble phosphorus are about three times as high, which is not an unreasonable difference, since both the seasonal and the annual variations might well be large. It is more difficult to explain the great discrepancy in the data for combined P, since their values are over one hundred times as great as mine. But regardless of which estimate is more nearly correct, it does not alter the general conclusion that the situation is a different one from that described by Atkins in the Plymouth Sound region. The phosphates in the Mississippi are not exhausted before they reach the mouth of the river but are poured into the Gulf in large quantities, in a form readily utilized by marine phytoplankton.

The same is probably true of nitrogen compounds. No analyses are available either of Gulf waters or at the mouth of the river, but analyses at New Orleans indicate that the quantities are rather large. I am indebted to Mr. Hirsch for the following data, which are approximate averages for 1935:

Ammonia N	20 mg. per m^3.
Albuminoid N	350 mg. per m^3.
Nitrite N	5 mg. per m^3.
Nitrate N	200 mg. per m^3.

The average nitrate concentration in the river is thus approximately the same as the accumulated nitrate concentration in the depths of the ocean but much greater than the normal concentration in the marine surface layers into which the river outflow mixes itself. The ammonia and albuminoid nitrogen contents of the river (assuming that the latter is decomposed sooner or later with the liberation of NH_3) are also many times as great as those normal to the marine surface layer.

The effect of the Mississippi on the Gulf

It has been shown that the Mississippi releases a large quantity of phosphate into the Gulf. From two-thirds to nine-tenths of it is in soluble form, capable of being utilized by marine plants as soon as the river water becomes mixed with Gulf water. The amount of soluble phosphate per unit volume in the river is approximately four times the mean amount found in the sea water of the surrounding region. It might therefore be expected to exert a considerable biological influence provided the amount of flow from the river is great enough to cause a significant increase in the phosphate content of the surrounding waters.* That such an increase does take place is indicated by fig. 21.

It is assumed that this increase is due largely to mixing with the river water because: 1. The low salinity of this water proves that such mixing

* See footnote on p. 68.

has occurred. 2. There is little evidence of an important increase by other means, such as regeneration from the mud by vertical mixing or upwelling of deep water. The greatest concentration of phosphates is found in the low salinity surface layer, with lesser concentrations at depths of ten to twenty-five meters. But regeneration from the bottom would suggest a greater concentration at the lower levels, and the presence of a distinct salinity gradient is ample evidence that vertical convection had not occurred to any great extent.

In the same general region phytoplankton production appears to be high. As may be seen by comparison of figures 21 and 22, the area of high chlorophyll content coincides with that of high phosphorus except that the former extends somewhat farther west and not so far south.

It is obvious that the amount of chlorophyll per unit volume is greater in the waters near the Mississippi. It appears also that the absolute amount is greater. This may be demonstrated in the following way: Station 2829 was chosen as a typical example for the region of high chlorophyll content, and the most useful one available because of the completeness of its data. A graph was drawn of the chlorophyll distribution down to 35 m., which was probably about the compensation point, since the chlorophyll was fairly low in spite of a high phosphate content. The curve was integrated planimetrically, and it was found that a column of water extending to a depth of 35 m. and 1 cm.^{2} in cross section would contain 23.2 Harvey units of chlorophyll. Farther out in the Gulf, at station 2838, the chlorophyll measured 2.6 units at the surface and 1.8 at 20 m. At this station the amount of chlorophyll per cm.^{2} in the first 20 m. was 4.4 units. If we assume that the water contained a uniform concentration of 1.8 units from 20 m. down to an indefinite depth, then the compensation point would have to be at a depth of 120 m. in order to make the photosynthetically active chlorophyll content as great as at station 2829. At 2824, a similar deep water station, it would have to be at 135 m. These values are impossibly deep, and we may conclude that the absolute amount of chlorophyll is greater in the region around the mouth of the Mississippi except in the very shallow water near shore.

A higher standing crop of phytoplankton in the Mississippi region means that the productivity of that area is greater provided there are no significant differences in the rates of phytoplankton growth and of consumption by zooplankton. There is a more plentiful supply of nutrients near the river to balance against the slightly higher temperature of the offshore waters. The rate of consumption appears to be more important from the standpoint of productivity than the growth rate, and such evidence as is available (ratios of total organic matter in the plankton to chlorophyll and the indirect evidence of the distribution of larger animals in the Gulf) indicates a larger quantity of zooplankton near the Mississippi. It may therefore be

concluded tentatively that the productivity is greatest in the high chlorophyll zone, although the matter cannot be settled definitely until more adequate data are available on the rate of plankton turnover in the Gulf.

Since the assumption is warranted that an increase of phosphorus content leads to greater phytoplankton productivity provided its growth is not limited or conditioned by other factors, the conclusion is permissible that the zone of high phosphate, and consequently the Mississippi itself, is an important factor in the productive area around the mouth of the river.

It was pointed out above that the zones of high phosphorus and high chlorophyll do not exactly coincide. The area where there is a high direct relationship lies within a radius of thirty or forty miles from the mouth of the river except toward the west, where it extends about eighty miles. Outside this region the relationship is slight, and consequently the correlation between phosphorus and chlorophyll in the entire area studied is low ($r = .10 \pm .05$). On purely theoretical grounds a low positive correlation or a correlation of zero is indicative of intrusion of phosphate into the area, since if the spring season begins with a more or less uniform concentration throughout the entire region, that part which has the greatest growth of phytoplankton will use the phosphorus most rapidly, thus leading to a negative relationship. This confirms the evidence presented above on the role of the Mississippi. But together with the fact that at none of the stations examined was there complete exhaustion of phosphate, it precludes the possibility that phosphorus is a limiting factor in these waters.

There are several possible explanations in regard to limiting factors:

1. Some other chemical substance or substances are involved.

2. The limiting factor is consumption by animals, which is rapid enough to keep the phytoplankton population below the level of complete utilization of phosphate.

3. During the spring season there are no limiting factors in the strict sense, but the greater quantity of nutrients near the mouth of the river conditions a higher plant growth there.

Figure 25 throws valuable light on the matter. Between salinities of 30.00 and 36.10‰ there is an inverse relation between salinity and chlorophyll with a coefficient of correlation of $-.68 \pm .16$. Below 30‰ the rate of increase of phytoplankton with decreasing salinity declines considerably. The data are too meagre for the remainder of the curve to be statistically significant, but there may be a secondary minimum at a salinity of about 10‰. The depression of phytoplankton growth in the lower salinities is partly due to turbidity. But the latter becomes apparent at the same point where the ratio of organic matter to chlorophyll begins to rise abruptly, that is, at a salinity of 18‰, whereas the depression of the chlorophyll curve is apparent at 30‰. Furthermore the inhibiting effect is not so marked in the Mississippi itself, where obviously the effect of turbidity is

greatest. Therefore, another factor is involved, probably salinity itself, which, in the brackish water zone, has an inhibiting effect on both marine and freshwater plankton.

The evidence just presented shows a definite relationship between chloro-

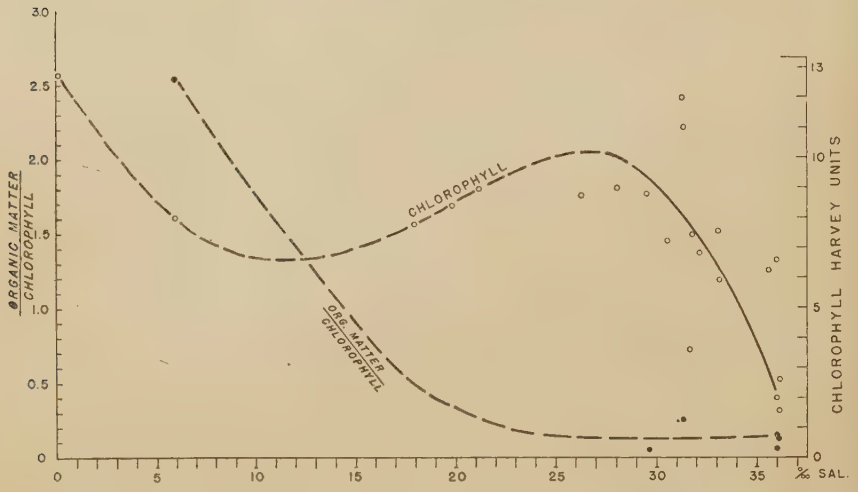

Figure 25. Relations of chlorophyll to organic matter and to salinities.

phyll and salinity. Up to a certain point, mixing of river water with the Gulf causes an increase in the phytoplankton crop. It is therefore probable that there is some nutrient salt in the river water, occurring there in higher concentration than in the Gulf and diffusing into it in much the same manner as phosphorus, which is a limiting factor in this region.

SUMMARY

1. The Mississippi River carries large quantities of nutrient salts into the Gulf of Mexico. The quantity of phosphate per unit volume is approximately four times as much as the amount found in the sea water of the surrounding region.

2. In the Gulf there is a superficial zone of high phosphorus content around the mouth of the Mississippi, extending from Mobile Bay westward to longitude 91°. Evidence is presented to show that this phosphorus is largely derived from the Mississippi and that the soluble phosphate of the river is more important in this respect than that in organic combination.

3. There is a zone of high phytoplankton content near the river which roughly coincides with the area of high phosphorus. A causal connection between the two is indicated.

4. It is suggested on purely theoretical grounds that some other nutrient which is discharged from the Mississippi in the same manner as phosphate is a limiting factor in this region.

5. A few notes are presented on the distribution of copper in the Gulf. It was present in all samples analyzed in both soluble and combined form. In deep water there appears to be a slight accumulation of soluble copper in the hypothalassa.

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TABLE OF OBSERVATIONS

Station*	Depth meters	Salinity ‰	Chlorophyll Harvey units per liter	Sol. Cu. mg. per m^3	Sol. P. mg. per m^3	Organic Matter mg. per liter	Seston Cu. mg. per m^3
A	1	31.47	11.1	9	11	2.9	3.2
	6	31.24	7.2	4	15	4.0	5.4
2812	1	31.70	3.6	9	8		2.3
	15	35.52	5.8	9	5		2.6
2815	1	29.62	8.8	25	4	.44	1.3
	25	36.08	5.0	20	4	.14	1.6
2817	1	31.56	6.2	11	3		
	25	36.17	2.5	8	2		
2821	1	18.01	7.8	12	16	4.27	6.8
	13	34.08	4.8	12	11	1.00	1.5
2824	1		2.0	5	4		
	25		1.7	8	3		
2827	1	5.95	8.1	15	16	20.60	7.2
2829	1	19.90	8.4	14	7	.49	.8
	10	26.40	10.2	7	2		
	20		4.8	7	2	.32	.3
	35		4.5	5	10	.81	.7
2832	1	36.05	1.6	7	4	.23	.5
2835	1	35.91		5			
	100	36.00		9			
	300			10			
	600			10			
	1800			12			
2838	1	36.10	2.6	3	6	.34	.4
	20	36.09	1.8	3	4		
2839	1	26.82	8.7	8	2		
	10	32.55	4.8	6	5		
2844	1	30.57	7.2	6	2		
2847	1	36.00	6.4	10	2	.38	.7
B	0	31.35	12.0	2	3		
C	0	31.85	7.4	4	3		
D	0	33.19	5.9	1	4		
E	0	33.17	7.5	2	2		
F	0	32.22	6.8	3	3		
G	0	28.03	9.0	2	2		
H	0	21.24	9.0	2	3		
Mississippi River New Orleans, La. April 5, 1937			12.8	15	6		

Seston Phosphorus
Station 2827. 1.6 mg./m^3 at 1 meter; Station 2829. 0.5 mg./m^3 at 10 meters; Station 2838. 0.3 mg./m^3 at 20 meters.

* See figure, page 61.

THE SEX-RATIO IN ALTERNATIONAL HERMAPHRODITES, WITH ESPECIAL REFERENCE TO THE DETERMINATION OF RATE OF REVERSAL OF SEXUAL PHASE IN OVIPAROUS OYSTERS

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In a population in which the sexual nature of the individuals is not subject to change, the sex-ratio of a given brood as determinable in practice may be different at different times if there is a differential rate of maturation for the two sexes, or a differential death rate. Excluding from present consideration the periods in the life of the population at which these differentials may operate, however, the sex-ratio of the brood at any given time should be identical with that at any other time. In a population of alternational hermaphrodites, on the other hand, in which the sexual nature of the individuals is subject to change, the sex-ratio should at any given time be a function of the initial sex-ratio and the rate of reversal of sexual phase. It is obvious that if the process of reversal is continuous, this function will be one of higher power. In many organisms, however, the process is expressed discontinuously, in separated integrals related to discontinuous sexual activity, and the sex-ratio of a population of such alternational hermaphrodites in any given spawning period may accordingly be expressed in first approximation as a rectilinear function of the sex-ratio of the immediately preceding period of sexual activity and the relative number of individuals in which reversal of sexual phase has occurred, according to the equation:

$$M_{n+1} = M_n + F_n\delta_{Fn} - M_n\delta_{Mn} \qquad (1)$$

This equation might be stated in words to the effect that the males of a given brood at a given spawning period (employing this term in the sense of the sexually active portion of the reproductive cycle of the population) must be equal in number to the males of the preceding spawning period, plus the females which have in the interval become males, minus the males which have in the interval become females. The symbols employed in the equation may be precisely defined as follows:

M_{n+1} = male population in the spawning period under consideration, expressed as a fraction of the total sexually active population at the time (a mode of expression justified by the fact that, if it be assumed for present purposes that the rates of mortality or of maturation are not differential, the male population changes only as does its ratio to the total population).

M_n = male population in the spawning period preceding the one under consideration, expressed as a fraction of the total sexually active population at that time.

F_n = female population in the preceding period, expressed as a fraction of the sexually active population at that time (= 1.00 − M_n, the reciprocal of M_n).

δ_{Mn} = males of the preceding spawning period which are functioning as females in the period under consideration, expressed as a fraction of the population of males in the preceding period.

δ_{Fn} = females of the preceding period which have been converted into males, expressed as a fraction of the population of females from which these new males were derived.

M_1 would refer to the male fraction of the sexually active population at the first spawning period of the brood. δ_{M1} would refer to the fraction of M_1 functioning as females at the second spawning period (when these individuals, formerly a part of M_1, would constitute a part of F_2).

The expression of the sexual changes of alternational hermaphrodites in terms of amount of reversal of phase permits a comparison of sexual behaviour in different age-groups, localities, or species which cannot be made by simple examination of the sex-ratios of successive years, as will be emphasized in further paragraphs. The sex-ratios are, however, readily obtained,* whereas δ_{Mn} and δ_{Fn} on the other hand can be directly determined only by an experimental study extending from spawning-period (n) through spawning-period ($n + 1$). It thus becomes desirable to employ equation (1) for determining δ_{Mn} and δ_{Fn} from known M_n and M_{n+1}; and it is therefore convenient to derive from it the following equations:

$$\delta_{Mn} = \frac{M_n - M_{n+1} + F_n\delta_{Fn}}{M_n} \tag{2}$$

$$\delta_{Fn} = \frac{M_{n+1} - M_n + M_n\delta_{Mn}}{F_n} \tag{3}$$

Given the sex-ratios only, it is obvious that the amount of reversal of sexual phase in one sex cannot be obtained in absolute but only in relative values, in terms of the amount of reversal in the other sex; since it cannot be determined without experimental study whether, for instance, certain increases in the proportion of males from one spawning-period to the next have resulted from a relatively low rate of reversal in males or a relatively high rate in females. Absolute limiting values may, however, be obtained for δ_{Mn} and δ_{Fn} by introducing for one of them its extreme limiting values 0 or 1.0 (which are set by the fact that δ_{Mn}, etc., represent the ratios of a part to the whole of a population). The circumstances under which, when one of the variables δ_{Mn}, δ_{Fn} is set equal to its extreme limits, minima or

* Indeed, as the sex-ratio at a given spawning period seems in certain alternational hermaphrodites to be more or less constant from brood to brood in a given environment, it is sometimes even possible in practise to obtain approximate values for M_n and M_{n+1} by simultaneous determination of the sex-ratios of successive year-classes of a local population.

maxima respectively greater than 0 or less than 1.0 are obtained for the other, may be determined from equations (2) and (3) in the following manner: Let δ_{Fn} in (2) equal its extreme maximum value of 1.0; then δ_{Mn} will obviously be at its maximum, and will have the following value:

$$\delta_{Mn\ \max.} = \frac{M_n - M_{n+1} + F_n}{M_n} = \frac{1.0 - M_{n+1}}{M_n} = \frac{F_{n+1}}{M_n} \tag{4}$$

According to (4), δ_{Mn} max. will be equal to or less than 1.0 when F_{n+1} is equal to or less than M_n. When F_{n+1} is greater than M_n (which would result in δ_{Mn} max. > 1.0), the equation yields no practical answer to the question of maximum value of δ_{Mn} other than to indicate that it may attain the extreme maximum of 1.0 to which it is limited by definition.

By means of operations analogous to the above [setting δ_{Fn} equal to its extreme minimum, 0, in equation (2), and δ_{Mn} equal to its extreme limits in equation (3)], the equations may be made to yield absolute limiting values of practical significance—that is, numerical limits for either variable within its limits by definition (0 and 1.0)—in the following circumstances:

(*a*) When $M_{n+1} < M_n$, $\delta_{Mn\ \min.} = 1.0 - \frac{M_{n+1}}{M_n} = > 0$

(*b*) When $M_{n+1} > M_n$, $\delta_{Fn\ \min.} = 1.0 - \frac{F_{n+1}}{F_n} = > 0$

(*c*) When $M_{n+1} < F_n$, $\delta_{Fn\ \max.} = \frac{M_{n+1}}{F_n} = < 1.0$

(*d*) When $F_{n+1} < M_n$, $\delta_{Mn\ \max.} = \frac{F_{n+1}}{M_n} = < 1.0$

When any circumstances other than the above prevail, the minima and maxima may attain the extreme limiting values 0 *and* 1.0.

To check the above general limiting values against a specific case, let for example:

$M_n = M_1 = 0.25$ (and therefore its reciprocal $F_1 = 0.75$)

$M_{n+1} = M_2 = 0.50$ (and therefore its reciprocal $F_2 = 0.50$)

Substituting these values in equation (3) we obtain:

$$\delta_{F_1} = \frac{0.50 - 0.25}{0.75} + \frac{0.25}{0.75}\delta_{M_1} = 0.33\delta_{M_1} + 0.33 \tag{5}$$

Inasmuch as the given M_{n+1} is greater than the given M_n, and less than the given F_n, the conditions presented correspond to those under (*b*) and (*c*); and accordingly δ_{F_1} max. = 0.66 and δ_{F_1} min. = 0.33, and the proportion

of females which have been converted into males since the preceding spawning period must lie between these numerical limits. Since the conditions do not correspond to (*a*) or (*d*), it appears that that δ_{M_l} can reach (or lie anywhere between) the extreme limits 1.0 and 0, dependent on the actual value of δ_{F_l}.

If inferences from results yielded by substitution of actual data in the equations given above are to be considered as valid, the inferences and the data must satisfy the following conditions:

I. Unless both sexes are known to be subject to reversal of sexual phase, it cannot be inferred that the rates of reversal may be other than the least absolute values permitted by the equation. Since minimum values other than 0 for δM_n are only obtained when $M_{n+1} < M_n$ and for δ_{Fn} when $M_{n+1} > M_n$, and since these requirements are antithetic and cannot be simultaneously fulfilled, it follows that reversal of sexual phase in *both* sexes can be inferred from the sex-ratios only when the value of the term $(M_{n+1} - M_n)$ of equation (3) changes its sign in the course of the history of a brood. Thus, in the example (5), unless it is given that the organism for which the sex-ratios are assumed is more than strictly proterogynous, the inference that δ_{Fn} may have a value greater than the absolute minimum, 0.33, is unjustified, even though the equation indicates a possible maximum of 0.66.

II. Unless reversal of sexual phase is known to be discontinuously exhibited in the population, as from one spawning period to the next in a given brood, the relative rates of reversal indicated by the equation cannot be inferred to represent other than a residual fraction of the actual rate. Thus, an equation of higher power would be required for determining rate of reversal in the viviparous oysters, among which it appears that reversal of sexual phase may occur in an individual several times during the annual spawning-period of the brood.

III. Unless the rate of maturation or of mortality is known to be the same in both sexes, at least during that portion of the life of the brood at which sex-ratios considered are obtained, values of M_{n+1}, etc. substituted in the equation must have been corrected for differences in these rates.

IV. Unless it is known that the sex-ratios at a given age are constant from brood to brood in the locality—or environment—considered, the changes in ratio must have been actually followed in a single brood from spawning-period to spawning-period.

Among alternational hermaphrodites in which the sexual dynamics might seemingly be studied to increased advantage by the method here suggested, various observations as to both the sex-ratios and the absolute amounts of reversal of phase are available for the oviparous oysters. Although the data are entirely too meager to provide conclusive evidence, it

seems of interest to present the results obtained from their substitution in the equations, both as a demonstration of the practical value of these equations and because the results expose certain interesting possibilities not revealed by previous methods of comparison of the sex-ratios. Thus, there may be noticed consistent if tenuous indications of a qualitative correlation between the initial sex-ratio and the subsequent rates of reversal of sexual phase, from which, since the initial sex-ratio according to all observers seems to be influenced by environment, it might be inferred that the subsequent reversals may be subject to this same affect. There are also indicated to be changes in rate of reversal with age, and differences between species, which may imply the activity of other factors in addition to the environmental ones in determining reversal of phase in the individual. Finally, the results of the calculations suggest a possible quantitive correspondence between initial sex-ratio and initial rate of reversal of sexual phase (for which, however, the data available are very inadequate indeed). Should more extensive information confirm this latter relation, it would seem necessary to question whether the entire population of the oviparous oysters may not, prior to its initial spawning, attain a state of maleness which under certain environmental conditions may give way to femaleness before reaching functional expression, in a degree analogous with the degree of conversion of functional males into females by the time of the second spawning period.

Some doubt seems to have been felt in various quarters as to the interpretation of observations by Amemiya, 1929, and by Needler, 1932a and b, to the effect that individuals of oviparous oysters functionally female may exhibit maleness in a succeeding year, and the converse; especially as to whether the observed high proportions of changes were normal to the species, rather than artifacts of the operative procedure previously employed in determining the sex of experimental material. Therefore, before presenting the results of conversion into relative rates of reversal of the sex-ratios derived from oviparous oysters, fresh experimental information [the by-product of an examination of methods for analysis of the effect of grouping upon sexuality (*cf* Burkenroad, 1931)] may be set forth to indicate that a considerable proportion of reversal normally occurs among both sexes of these animals, as suggested by the sex-ratios.

Sixty-six individuals among a stock of *Ostrea virginica* from the beds about Milford, Connecticut, were induced to spawn by subjection to temperatures varying between 26° and 34°, modally 30° C; and their functional sex in this manner determined. No deaths were observed among either the spawners or the stock from which they were separated during the period of determination, July 28 to August 22, 1934. The oysters which had been determined as to sex were identified by shell-cuts and set out on good ground at a depth of twenty feet, in partitioned trays the tops and bottoms of which were wire mesh and which were supported with six-inch clearance above concrete anchor-blocks. One to six oysters, in various combinations of age and of sex or in the company of round clams, were placed in each compartment of the trays. When examined the following year, August, 1935, only four deaths had occurred, and both oysters and clams had grown vigorously. An unfortunate arrangement for staorge

REVERSAL OF SEX IN *O. virginica*

Age, 1934	*Functional Sex, 1934*		*Functional Sex, 1935*			
	♂	♀	♂	♀	☿	*Indet.*
2 years	10	0	6	3	0	1
4 years	17	0	10	4	1	2
2 years	0	7	2	3	0	2
4 years	0	9	2	5	0	2

of the recovered material preparatory to redetermination of sex by spawning resulted in complete loss of the contents of one tray and injury to the forty-three remaining individuals, which were therefore fixed and examined histologically, with the results shown in Table I. The number of individuals available, while sufficient to establish that reversal of sexual phase occurs so frequently in *O. virginica* of advanced age as perhaps even to indicate for this species an alternating rhythm comparable with, although of much slower rate than, that of viviparous oysters, is far too small for detailed analysis. The anomalous result, that both the retention and the acquisition of femaleness were greatest in the more crowded compartments, may however be mentioned.

Accepting that the oviparous oysters are actually alternational hermaphrodites; and other conditions stated on p. 78 being assumed to be fulfilled within the limits required for present purposes,* the M_1 and M_2 found by Amemiya (1928) among "one-" and "two-winter" individuals of the Japanese oviparous oyster, *O. gigas*, on "Ground A," may be calculated and substituted in equation (3), with the following result:

$$\delta_{F_1} = 0.941\delta_{M_1} - 0.126 \qquad (6)$$

In an experiment by Amemiya (1929) with "one-winter" oysters (presumably of "Ground A" although this is not explicitly stated to be so), 49% of the surviving and determinable individuals which had been males in the preceding year were found to be functioning as females at their second spawning. If this value of δ_{M_1} is substituted in equation (6), a calculated δ_{F_1} of 0.336 is obtained, which is in close agreement with Amemiya's experimentally determined δ_{F_1} of 0.32. Since the oysters from which Amemiya obtained his data on the sex-ratio were born a year before those in which he examined the amount of reversal of sexual phase, this correspondence between the observed and the calculated percentages of reversal, if not a coincidence, would suggest that δ_{M_1} and δ_{F_1} may tend to be constant from brood to brood in a given locality.

From the sex-ratios found by Amemiya among 1^+ and 2^+ *O. gigas* at "Ground B," it may be calculated that

$$\delta_{F_1} = 1.16\delta_{M_1} + 0.154 \qquad (7)$$

On comparison of this with the equation (6) for "Ground A" it appears that the proportion of reversal differs from one of these localities to the other, necessarily in one sex [as is seen if the value of δ_{M_1} experimentally determined at "Ground A," 0.49, is substituted in equation (7)] and probably in

* See the discussion of recent data supplied by Needler, on p. 82.

both. It also appears that, whereas at "Ground A" the proportion of females converted into males (δ_{F_1}) must be less than the proportion of males converted into females (δ_{M_1}), at "Ground B" the reverse is the case.

Coe (1932, 1934) has shown that F_1 is locally variable in *O. virginica*, but is annually rather constant in a given locality. Amemiya's data dealt with in the preceding paragraph show a large difference in F_1 between "Ground A" and "Ground B" in *O. gigas*. If it be assumed that in agreement with *O. virginica* the initial sex-ratios in *O. gigas* are locally constant from brood to brood, a possible similarity in the behaviour of the initial sex-ratio to that of the rates of reversal of sexual phase is suggested, since the formulations in the preceding paragraph indicate the possibility of variation with locality, but constancy in successive broods in the same locality, of the rate of reversal in *O. gigas*. Further, the indication in the preceding paragraph that not only is the M_1 greater at "Ground B" than at "Ground A," but the rates of sex-reversal also more favorable to maleness at the former locality, suggests that rates of reversal may perhaps vary not only in the same manner as do initial sex-ratios, but in the same direction, and therefore that both may be under the influence of the same affect. It is perhaps worthy of note that the initial proportion of females at "Ground A," 0.52, is rather close to the rate of reversal of males (δ_{M_1}) observed by Amemiya, 0.49. If such a correspondence were to be true for "Ground B," where the initial proportion of females is 0.46, then the rate of reversal of females (δ_{F_1}) at this bed must have a value of something like 0.69, as compared with 0.32 at "Ground A." It would be interesting to know if this is the case, in view of the possibility that F_1, in populations of oviparous oysters where the first spawning occurs only after the first winter, is the result of a hypothetical δ_{M_0} in a hypothetical M_0 of 100% [an M_0 which might conceivably become an actual M_1 of 1.00 (complete protandry) in an environment permitting functional expression of sexuality before the process of reversal becomes effective.*]

In *O. virginica* the F_1 [average value 0.11, records of Coe (1934) and of Needler (1932a) combined] is so low as compared with *O. gigas* that a very small δ_{M_1} would be sufficient to effect the observed considerable increase in proportion of females at the second spawning season (to an average F_2 of 0.25). As an average in *O. virginica*

$$\delta_{F_1} = 8.45\delta_{M_1} - 1.34 \qquad (8)$$

so that the maximum possible δ_{M_1} (at the limiting value $\delta_{F_1} = 1.0$) would be only 0.28. By contrast, a maximum δ_{M_1} of 0.72 would be possible in *O. gigas* at "Ground B," while at "Ground A" the maximum could reach the limiting value, 1.0 (although its experimentally determined actual value is

* Cf. pp. 82–83.

only 0.49). The high initial proportion of females in *O. gigas* (an average F_1 of 0.39 for Amemiya's two localities) has been expressed by Amemiya as an absence of protandry, but there seems no reason to believe that it is the result of other than the operation with a different value of δ_{Mo} of a process homologous with that in *O. virginica,* where the functional results (average $F_1 = 0.11$) are described by Coe as a strong tendency to protandry. The fact that in *O. virginica* an F_1 lower than in *O. gigas* is correlated with a δ_{M_1} also lower than in the Japanese species [according to the low maximum indicated by equation (8)] suggests that perhaps between species of oviparous oysters (as analogously, between local populations within a species) the same factors that are responsible for differences between the species in their initial sex-ratios also determine similar differences in the rates of reversal of sexual phase.

From the changes in average sex-ratio in *O. virginica* between the second and third and between the third and fourth years indicated by the records of Needler (1932a), it may be calculated that

$$\delta_{F_3} = 3.05\delta_{M_3} - 0.765 \tag{9}$$

$$\delta_{F_3} = 1.29\delta_{M_3} - 0.378 \tag{10}$$

which suggests that the rates of sex-reversal may tend to become more favorable to femaleness with advance in age. According to these equations, either the δ_{Mn} indicated by the sparse data provided by Needler's experiments and those reported above is too low in relation to the observed δ_{Fn}, or the latter is too high.

Since the completion of the foregoing, some observations of changes in sex-ratio from year to year among *O. virginica* of the same brood have been published by Needler (1936). Calculating from her graph the percentages of males and females among those oysters in which sex was determinable, values for δ_{Fn} comparable with those given in preceding paragraphs (which were derived from the averages of a number of localities) are obtained, namely:

$$\delta_{F_1} = 8.51\delta_{M_1} - 1.73; \quad \delta_{F_2} = 2.48\delta_{M_2} - 0.93; \quad \delta_{F_3} = 0.79\delta_{M_3} - 0.19.$$

These figures seem to justify the assumption on p. 80 above that the data then available for *O. virginica* were acceptable for the uses intended.

In another pertinent and most valuable paper appearing since the elaboration of the foregoing account, the suggestion made above (that there may be an essentially complete protandry in *O. virginica* although one not necessarily attaining functional expression) has been tentatively accepted by Coe (Biol. Bull., LXXI, 2, 1936; pp. 357–8); who at the same time, however, repeats without qualification the generalization made in his previous works, that "The more favorable the conditions are for the rapid growth of the young oyster, the more likely is the initial male phase to be

omitted or aborted." Since the hypothesis of complete protandry permits an alternative interpretation of the facts to which Coe's generalization refers, it seems desirable to describe this hypothesis and its corollaries in somewhat greater detail than on p. 81 above. If the relation of environment to sexuality in oviparous oysters should be viewed simply as one of rate of reaction, environment might be regarded as exercising a quantitative effect upon the number of cycles of alternation of sexuality which may be completed within a given period, but as producing a qualitative effect upon functional sexuality only to whatever extent the physiological state observed as predominance of maleness is capable of transformation to that of femaleness before it has attained the level of functional expression. If it should be assumed that a certain level (but not necessarily a functional one) of male predominance or development must be attained by the young oysters before predominance of the female state can occur, then several types of environmental affect of the sex-ratio at first spawning might be distinguished, such as that by which sexual development during the first summer of life would be (1) so late in inception, or so slow, as to be checked by the onset of winter before attainment of the required minimum grade of maleness, with resultant completion of male development during the second summer and a proportion of males at first spawning which might in ideal circumstances approach 100%, or complete functional protandry; or (2) sufficiently early or rapid as to reach the minimum grade of maleness, but not functional sexuality, before the first winter, so that while initial spawning occurs [as in (1)] during the second summer, the attained capacity for reversal of the aborted male phase results in an appreciable F_1, commensurate with what I have termed δ_{Mo}; or (3) so early or rapid as to terminate in functional maleness during the first summer, with a consequent approach to complete protandry; in which circumstance the proportion of females at first spawning would have been diminished rather than increased by a more favorable environment. It is possible that conditions to some extent approaching those outlined in (3) distinguish *O. virginica* in the warm waters of Louisiana [*cf.* McConnell, J. N., 1931. The Louisiana Oyster. Bull. Dept. Cons. Louisiana, I (revised), p. 23. "Recent work done by M. D. Burkenroad . . . indicates that . . . the oyster usually becomes a sexually mature male in the same spawning season in which it was born, about two or three months after it has settled . . . It is at this time about one-half to three-quarters of an inch long [body length]"].

SUMMARY

An equation descriptive of the changes in sex-ratio in organisms subject to discontinuously expressed reversal of sexual phase is discussed. Substitution of actual data in the equation reveals the possibility that rate of reversal of phase may be correlable with initial sex-ratio in the oviparous oyster.

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